1-1-2013

Evolution of a Cambrian active continental margin: the Delamerian-Lachlan connection in southeastern Australia from a zircon perspective

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Abstract

The Early Palaeozoic Ross–Delamerian orogenic belt is considered to have formed as an active margin facing the palaeo-Pacific Ocean with some island arc collisions, as in Tasmania (Australia) and Northern Victoria Land (Antarctica), followed by terminal deformation and cessation of active convergence. On the Cambrian eastern margin of Australia adjacent to the Delamerian Fold Belt, island arc and backarc basin crust was formed and is now preserved in the Lachlan Fold Belt and is consistent with a spatial link between the Delamerian and Lachlan orogens. The Delamerian–Lachlan connection is tested with new zircon data. Metamorphic zircons from a basic eclogite sample from the Franklin Metamorphic Complex in the Tyennan region of central Tasmania have rare earth element signatures showing that eclogite metamorphism occurred at ~ 510 Ma, consistent with island arc–passive margin collision during the Delamerian(– Tyennan) Orogeny. U–Pb ages of detrital zircons have been determined from two samples of Ordovician sandstones in the Lachlan Fold Belt at Melville Point (south coast of New South Wales) and the Howqua River (western Tabberabbera Zone of eastern Victoria). These rocks were chosen because they are the first major clastic influx at the base of the Ordovician 'Bengal-fan' scale turbidite pile. The samples show the same prominent peaks as previously found elsewhere (600–500 Ma Pacific-Gondwana and the 1300–1000 Ma Grenville–Gondwana signatures) reflecting supercontinent formation. We highlight the presence of ~ 500 Ma non-rounded, simple zircons indicating clastic input most likely from igneous rocks formed during the Delamerian and Ross Orogenies. We consider that the most probable source of the Ordovician turbidites was in East Antarctica adjacent to the Ross Orogen rather than reflecting long distance transport from the Transgondwanan Supermountain (i.e. East African Orogen). Together with other provenance indicators such as detrital mica ages, this is a confirmation of the Delamerian–Lachlan connection.

Keywords

active, continental, margin, delamerian, lachlan, cambrian, connection, evolution, southeastern, australia, zircon, perspective

Disciplines

Medicine and Health Sciences | Social and Behavioral Sciences

Publication Details

Fergusson, C. L., Nutman, A. P., Kamiichi, T. & Hidaka, H. (2013). Evolution of a Cambrian active continental margin: the Delamerian-Lachlan connection in southeastern Australia from a zircon perspective. Gondwana Research, 24 (3-4), 1051-1066.

Evolution of a Cambrian active continental margin: the Delamerian–Lachlan connection in southeastern Australia from a zircon perspective

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ABSTRACT

The Early Palaeozoic Ross–Delamerian orogenic belt is considered to have formed as an active margin facing the paleo-Pacific Ocean with some island arc collisions, as in Tasmania (Australia) and Northern Victoria Land (Antarctica), followed by terminal deformation and cessation of active convergence. On the Cambrian eastern margin of Australia adjacent to the Delamerian Fold Belt, island arc and backarc basin crust was formed and is now preserved in the Lachlan Fold Belt and is consistent with a spatial link between the Delamerian and Lachlan orogens. The Delamerian–Lachlan connection is tested with new zircon data. Metamorphic zircons from a basic eclogite sample from the Franklin Metamorphic Complex in the Tyennan region of central Tasmania have rare earth element signatures showing that eclogite metamorphism occurred at ~510 Ma, consistent with island arc – passive margin collision during the Delamerian(–Tyennan) Orogeny. U–Pb ages of detrital zircons have been determined from two samples of Ordovician sandstones in the Lachlan Fold Belt at Melville Point (south coast of New South Wales) and the Howqua River (western Tabberabbera Zone of eastern Victoria). These rocks were chosen because they are the first major clastic influx at the base of the Ordovician 'Bengal-fan' scale turbidite pile. The samples show the same prominent peaks as previously found elsewhere (600–500 Ma Pacific–Gondwana and the 1300–1000 Ma Grenville–Gondwana signatures) reflecting supercontinent formation. We highlight the presence of ~500 Ma non-rounded, simple zircons indicating clastic input most likely from igneous rocks formed during the Delamerian and Ross Orogenies. We consider that the most probable source of the Ordovician turbidites was in East Antarctica adjacent to the Ross Orogen rather than reflecting long distance transport from the Transgondwanan Supermountain (i.e. East African Orogen). Together with other provenance indicators such as detrital mica ages, this is a confirmation of the Delamerian–Lachlan connection.

Keywords:
Arc–continent collision
Convergent margin
Eclogite
Lachlan Fold Belt
Zircon

1. Introduction

An issue with wider implications for the development of the East Gondwana Pacific margin is the relationship between the Lachlan and Delamerian fold belts in southeastern Australia (Cawood, 2005; Foster et al., 2005; Squire and Wilson, 2005; Cayley, 2011). The long-held opinion has been that these two orogenic belts developed adjacent to each other with the Delamerian Fold Belt formed by Neoproterozoic continental rifting followed by orogenesis involving collision and convergent margin tectonism in the Delamerian(–Tyennan) Orogeny, whereas the Lachlan Fold Belt developed with Cambrian oceanic island arc and backarc basement followed by widespread Ordovician turbidite deposition and island arc development (Cas, 1983; Powell, 1984; Glen, 2005). The source of the Ordovician turbidites is clearly continental, as indicated by the quartz-rich nature of sandstones in the succession. Some authors infer derivation of the sandstones from the Delamerian mountain chain (Turner et al., 1996; Fergusson and Tye, 1999), whilst others argue for a more distant source in the East African Orogen at the junction of West and East Gondwana (Fig. 1) (the Gondwana Supermountain of Squire et al., 2006; Williams and Pulford, 2008). The East African Orogen has been considered a significant source for early Palaeozoic sedimentary units in Gondwana including North Africa and adjacent parts of the Middle East (Figure 1) (Squire et al., 2006; Meinhold et al., 2013) and may also have been a source for late Cambrian to earliest Ordovician sedimentary rocks along the Tethyan margin of north India (Cawood et al., 2007). It has also been argued that many of the early Palaeozoic units of the Lachlan Fold Belt are allochthonous and have been transported by hundreds of kilometres or more along strike-slip faults (Willman et al., 2002; Glen, 2005; Glen et al., 2009).

These interpretations can be tested by analysis and compilation of provenance data including paleocurrents, sedimentary petrology, and isotopic data including zircon and mica ages of detrital components (Cas, 1983; Ireland et al., 1998; Fergusson and Tye, 1999; Fergusson and Fanning, 2002; Squire et al., 2006; Williams and Pulford, 2008). Detrital zircon ages have been widely used to indicate provenance and for the Ordovician quartzose turbidites the detrital zircon age spectra are seemingly uniform with most ages in the ranges 600–500 Ma and 1300–1000 Ma. Similar spectra have also been widely found in early Palaeozoic sandstones throughout Gondwana (Veevers, 2000; Squire et al., 2006; Cawood et al., 2007; Williams and Pulford, 2008; Meinhold et al., 2013). In syntheses of global zircon ages these age ranges are associated with supercontinent formation. The 1300–1000 Ma range for granitoid ages and detrital zircons is associated with formation of Rodinia whereas the 600–500 Ma range reflects the final part of the Pan-African Orogeny during Gondwana assembly (Condie et al., 2009; Rino et al., 2008; Voice et al., 2011). Differences occur between detrital and igneous ages within continents; for example, no significant igneous peaks occur in Antarctica and Australia in the range 1500–1000 Ma whereas a large detrital peak is at 1060 Ma in Antarctica and at 1200 Ma in Australia (Condie et al., 2009, p. 237). How these are interpreted has important implications for provenance assessments. Thus does the absence of 1300–1000 Ma igneous ages in Antarctica reflect a sampling bias due to cover or is a distant source implied by the large detrital peak at 1060 Ma? Using detrital ages of zircons and other geochemical and isotopic data from sediment and sedimentary rocks it has been concluded that much of interior East Antarctica consists of an amalgam of Grenville and older cratons in a matrix of Pan-African fold belts (Veevers and Saeed, 2008, 2011).

We examine the connection between these orogenic belts by acquiring chemical and isotopic data from zircons in three widely spaced samples. The first sample is a ~510 Ma (Black et al., 1997) eclogite from the Franklin Metamorphic Complex of Tasmania and the other two samples are of sandstones from the base of the Ordovician turbidites of the Lachlan Fold Belt at the Howqua River of eastern Victoria and Melville Point on the south coast of New South Wales (Fig. 2). The study of the Tasmanian eclogite was undertaken to confirm if its zircons grew during collision-related eclogite facies metamorphism or whether they formed in some other event unrelated to high-pressure metamorphism. Although metamorphism at ~510 Ma has been shown by chemical U–Th–Pb monazite dating to be widespread in Tasmania (Berry et al., 2005) and therefore part of the Delamerian(–Tyennan) Orogeny, it has been assumed rather than demonstrated that the ~510 Ma SHRIMP U–Pb zircon age on eclogite in the Franklin Metamorphic Complex reflects the timing of high-pressure metamorphism. We have therefore undertaken rare earth element (REE) analysis of these zircons to see if they have a low HREE abundance and no negative Eu anomaly signature that is diagnostic of zircons grown during eclogite facies metamorphism (Rubatto, 2002). The study of the Ordovician turbidite samples was designed to provide a better constraint on the initiation and provenance of the extensive East Gondwana turbidite fan (Fergusson and Tye, 1999; Williams and Pulford, 2008). This was done by selecting the least mature Ordovician sandstones that occur at the base of the succession in the Howqua River in eastern Victoria and at Melville Point on the south coast of New South Wales. These samples are most likely to reflect the source in the earliest Ordovician, during the final phases of the Delamerian Orogeny (Foden et al., 2006).

2. Geological setting

2.1 Tasmanides of southeast Australia

The Tasmanides of southeast Australia consist of the inner Delamerian Fold Belt occurring in southeastern South Australia, western Victoria and western Tasmania and the outer Lachlan Fold Belt in New South Wales, Victoria and northeastern Tasmania (Fig. 2). The Delamerian Fold Belt consists of thick rift basin successions of Neoproterozoic age overlying Paleoproterozoic to Mesoproterozoic basement of the Gawler and Curnamona Cratons in South Australia (Preiss, 2000). In western Victoria, northwestern New South Wales and Tasmania, mafic volcanic rocks and associated sedimentary successions have been interpreted as part of a volcanic passive margin that formed at 600–580 Ma (Direen and Crawford, 2003a, b). In Tasmania, the Delamerian Fold Belt consists of Neoproterozoic basement and basinal successions overthrust by early Cambrian ophiolite complexes containing boninites indicative of early island arc development and succeeded by a middle to late Cambrian post-collisional volcanic and sedimentary succession (the Mt Read Volcanics) (Berry and Crawford, 1988; Crawford and Berry, 1992). The ophiolite complex was emplaced in an island arc – passive margin collision along an east-dipping subduction zone west of the island arc (Berry and Crawford, 1988). The basement Neoproterozoic quartzose siliciclastic successions contain abundant detrital zircons of 1800–1650 Ma with less abundant 1450–1400 Ma ones and some samples containing 1300–1000 Ma ages indicating a likely early Neoproterozoic age for the succession (Berry et al., 2001; Black et al., 2004). The position of Tasmania in the Tasmanides prior to the Ordovician is problematic with some authors arguing for its initial location along the East Antarctic passive margin prior to the Ross Orogeny (Berry et al., 2008; Gibson et al., 2011). In western Victoria, the Delamerian Fold Belt consists of the Glenelg and Grampians–Stavely zones that contain igneous and metamorphic rocks associated with the Delamerian Orogeny (Crawford et al., 2003). Boninitic and tholeiitic volcanic rocks have been documented in drill holes and the significance of widespread magnetic mafic rocks under cover has yet to be determined. They may represent thrust sheets, similar to those in Tasmania, that have been emplaced over the former late Neoproterozoic passive margin in an island arc – passive margin collision (Crawford et al., 2003). In western New South Wales, the Delamerian Fold Belt is preserved in the Koonenberry belt but in contrast to further south shows the development of a short-lived east-facing Delamerian active margin with an arc, forearc basin and eastern subduction complex (Greenfield et al., 2010, 2011), without an island arc–passive margin collision.

In contrast to the Delamerian Fold Belt, the Lachlan Fold Belt has an early to late Palaeozoic history. The tectonic development of the Lachlan Fold Belt is a widely debated issue. Numerous suggestions have been given for the Ordovician tectonic development including formation of multiple subduction zones (Gray and Foster, 2004), strike-slip emplacement of continental margin terranes (Glen et al., 2009) and island arc – passive margin collision somewhat akin to the Delamerian Orogeny in Tasmania (Aitchison and Buckman, 2012). In the Cambrian, an oceanic realm developed resembling the modern western Pacific Ocean with island arcs, backarc basins and boninitic-tholeiitic successions (Crawford et al., 2003). In contrast, the Melbourne Zone is underlain by a northward continuation of the basement rocks of western Tasmania called the 'Selwyn block' (Cayley et al., 2002; Cayley, 2011; Moore et al., 2013) as shown by magnetic connections across Bass Strait. Western Tasmania is thought to have reached its present location with respect to East Gondwana by the Early Ordovician (Li et al., 1997). In the earliest Ordovician, widespread quartzose turbidite deposition spread across much of the Lachlan Fold Belt following its earlier deposition in the Stawell Zone in the middle to late Cambrian (VandenBerg et al., 2000). Additionally, an island arc (Macquarie arc) was initiated in the east (Glen, 2005). The turbidite succession formed mainly through the Early to Middle Ordovician and in the eastern part of the fold belt was followed by widespread black shale and chert deposition (VandenBerg and Stewart, 1992).

The relationship between the Macquarie arc and the surrounding Ordovician quartz turbidites are problematic with several explanations offered including juxtaposition by strike-slip faulting (Glen et al., 2009), island arc rotation and development of adjoining subduction complexes (Fergusson, 2009), and overthrusting of a "passive margin" by an island arc from the east (Aitchison and Buckman, 2012). For the Early Ordovician, a connection between the Gondwana margin and the Macquarie Arc has been argued by Glen et al. (2011) who showed that the Mitchell

Breccia at the base of the Macquarie Arc succession contains a detrital zircon age pattern similar to that documented for the Ordovician quartz turbidites. The younger part of the Macquarie arc succession has a reduced number of detrital inherited zircons and it is unclear if this reflects a real trend or is due to an increasing proportion of magmatic zircons present in the samples (Glen et al., 2011, p. 682).

In the Late Ordovician widespread deformation, the Benambran Orogeny, caused crustal thickening in much of the central and eastern parts of the Lachlan Fold Belt into continental thicknesses such as in the Wagga–Omeo Zone with intrusion of abundant Silurian granites (Ickert and Williams, 2011). This new phase of magmatic activity has been related by Fergusson (2003, 2009) to migration of Macquarie arc magmatism westwards into the adjoining subduction complex. In the mid Silurian to Middle Devonian much of the Lachlan Fold Belt was in an extensional setting with widespread silicic igneous activity, thick successions of deep to shallow marine rocks and intermittent episodes of contractional deformation associated with an east-dipping subduction zone at the eastern margin of the belt (Powell, 1984; Collins, 2002).

2.2 Franklin Metamorphic Complex

The Tyennan block of central and southern Tasmania consists mainly of low-grade phyllite and quartzite with several higher grade complexes including the Franklin Metamorphic Complex. The Franklin Metamorphic Complex contains schist, minor quartzite, and lenses of metabasic rocks some of which are of eclogite grade (Meffre et al., 2000). These basic eclogites reached maximum metamorphic conditions at ~600–650°C and 1.5 GPa (Palmeri et al., 2009) and contain zircons with U–Pb ages of 511±8 Ma (Black et al., 1997; Turner et al., 1998). The schists have widely developed garnet and kyanite and therefore have most likely been metamorphosed at depth with the associated eclogites (Meffre et al., 2000).

2.3 Howqua River

The Howqua River crosses the contact between the Melbourne and Tabberabbera zones in eastern Victoria. The Melbourne Zone has a simply folded, upright Silurian to Early Devonian turbidite succession with a structurally complex belt further east (Mount Useful Fault Zone) that has windows containing middle to late Cambrian volcanic rocks of intermediate to silicic calc-alkaline affinity below a multiply deformed Late Ordovician black shale and Silurian turbidite succession (VandenBerg et al., 1995, 2006). Further east, the Tabberabbera Zone is dominated by an Ordovician and early Silurian turbidite succession overlain by Late Ordovician black shale (VandenBerg et al., 2000). In the Howqua River region, a largely complete Ordovician succession is developed with its basal earliest Ordovician turbidites underlain by bedded cherts which overlie a thick Cambrian altered mafic volcanic succession with upper tholeiitic and lower boninitic units (Fergusson, 1998; Crawford et al., 2003). Between the Cambrian volcanic rocks and the Mount Useful Fault Zone lies a structurally complicated assemblage of mélange containing chert, sandstone, mudstone and blueschist blocks with Late Ordovician ages based on poorly preserved graptolites in low-grade metasedimentary rocks, a U–Pb age on titanite and an Ar–Ar age on mica in blueschist (Spaggiari et al., 2002a). The Howqua River section is particularly significant as it is one of the few places in southeastern Australia where a basal contact to the widespread Ordovician turbidite succession is exposed, with reliably-dated Tremadocian–early Arenig graptolites (Fergusson, 1998). In contrast to the bulk of the overlying Ordovician turbidites, these basal sandstones have lower quartz (43–66%), higher feldspar (7–18%) and lithic clast (8–15%) contents (Fergusson and Tye, 1999).

The tectonic interpretation of the Howqua River region and the Lachlan Fold Belt is a matter of continuing controversy. In the mid 1990's, the idea of a double divergent subduction zone in central Victoria was introduced by Gray (1997) and Gray and Foster (1997) with subduction occurring towards the northeast generating the Tabberabbera Zone by subduction accretion of the thick turbidite succession and forming the Wagga–Omeo metamorphic belt in an arc setting. These ideas have been vigorously debated but more recent confirmation of the subduction-accretion related structural style of the Tabberabbera Zone has been provided by Watson and Gray (2001) and Willman et al. (2005).

2.4 Melville Point

Melville Point occurs on the south coast of New South Wales south of the southern end of the Permo–Triassic Sydney Basin. At Melville Point, the base of the Ordovician turbidites is exposed in a coastal cliff and rock platform (Fig. 2b). It is conformable on a succession of chert and mudstone that is underlain by altered basalt (Prendergast, 2007). The Ordovician turbidites are part of the Adaminaby Group whereas east of Melville Point the Wagonga Group is exposed. It is a package of Cambrian and Ordovician mafic volcanic rocks, limestone, quartz-rich turbidites, mudstone and mélange. Altered basalts, similar to those at Melville Point, are widespread in the Wagonga Group and have been shown to have an ocean island basalt magmatic affinity (Prendergast and Offler, 2012). Structures in the Wagonga Group are consistent with its interpretation as part of a subduction complex (Prendergast, 2007). The turbidites at Melville Point consist of thick, massive beds of quartz-feldspar-lithic sandstone interbedded with thin-bedded sandstone and mudstone containing thin chert layers. A layer of thin chert near the western end of the rock platform contains conodonts of late Cambrian to earliest Ordovician age (Glen et al., 2004, p. 873), thus showing that the underlying chert-mudstone and mafic volcanic rocks are of probable Cambrian age. Middle to late Cambrian limestones occur associated with mafic breccias located ~3.5 km east in the Wagonga Group at Burrewarra Point (Bischoff and Prendergast, 1987).

3. Zircon petrography, dating and rare earth element chemistry

3.1 Analytical methods

Zircon concentrates were prepared by standard heavy liquid and magnetic separation techniques at the mineral separation laboratory of the Research School of Earth Sciences, the Australian National University (ANU). The Howqua River and Melville Point sandstone zircon concentrates were hand picked under a binocular microscope, and the selected grains were cast in epoxy resin discs together with the zircon Temora 2 reference material (Black et al., 2003). The discs were ground to reveal mid-sections through the grains and then polished. The grains were documented with reflected and transmitted light photomicrographs and by cathodoluminescence (CL) imaging. CL images of representative zircons are shown in Fig. 3. The Franklin metamorphic zircons had been U–Pb dated in the 1990s (Black et al., 1997) and were located in the Geoscience Australia Mount Z1740, together with the U–Pb calibration zircon SL13. The mount was loaned by Geoscience Australia for additional CL-imagery guided U–Pb dating and REE analysis (Fig. 4).

Zircons were U–Pb dated on the SHRIMP II instruments at the University of Hiroshima and Geoscience Australia (Canberra). Analytical protocols followed Williams (1998), and reduction of the raw data used the ANU software 'PRAWN' and 'Llead'. $^{206}Pb/^{238}U$ of the Howqua River and Melville Point unknowns were calibrated using measurements of the Temora 2 reference material (U–Pb ages concordant at 417 Ma; Black et al., 2003). $^{206}Pb/^{238}U$ of the Franklin metamorphics zircons were calibrated using SL13 zircon. U and Th abundance was calibrated using measurement of the reference material SL13 (U = 238 ppm). The zircon U–Th–Pb data are summarised in Table 1. The reduced and calibrated data were assessed and plotted using the ISOPLOT program of Ludwig (2003).

Rare earth element analyses of some Franklin metamorphics zircons were undertaken using the Hiroshima University SHRIMP II instrument, in the normal mass-resolution mode (~6000), and according to the method outlined by Maas et al. (1992) and Hidaka et al. (2002). Abundances were calibrated using count rates obtained from analyses of the zircon standard 91500 (Wiedenbeck et al., 2004). Wiedenbeck et al. (2004) noted elemental heterogeneities within 91500. Given this observation, even if the very low abundances of La and Nd in our unknowns have been underestimated by an order of magnitude, this will not change the strongly fractionated, light REE depleted, heavy REE enriched (chondrite normalised) signatures of these zircons. REE analyses are summarised in Table 2.

3.2 Franklin metamorphics zircon morphology and U-Pb dating

Franklin metamorphics eclogite sample 9322–0004 yielded small (mostly <100 µm long) zircons of equant to stubby prismatic habit. In CL images their internal structure is complex, with domains of sector zoning, irregular patches and recrystallisation being more common than oscillatory zoning. A few grains have distinct narrow rims that appear homogeneous and bright in CL images (grain 8, Fig. 3). These features are typical of zircons grown during eclogite facies metamorphism (Nutman et al., 2008). Fourteen U–Pb analyses were undertaken on 9 grains (Table 1).

All sites have low Th/U (0.031 to 0.003), typical of metamorphic zircons. They all have close to concordant and statistically-indistinguishable U–Pb ages (Fig. 4). However, analysis 8.1 of a distinct rim overgrowth has a different REE pattern (see below) and thus was not included in the weighted mean $^{206}Pb/^{238}U$ age of 504±7 Ma (95% confidence, MSWD=0.41) for the other 13 analyses. These analyses included all varieties of sites within the grains (sector zoned, patchy and recrystallised). Anomalous site 8.1 on a narrow rim yielded a $^{206}Pb/^{238}U$ age of 460±35 Ma (1σ), whereas core analysis 8.1 gave an age of 510±14 Ma. Although these ages are within error of each other, the petrographic relationships indicate that the rim is younger.

3.3 Franklin metamorphics zircon REE chemistry

Five REE analyses were undertaken on the dated Franklin metamorphics eclogite sample 9322–0004 zircons: one on the distinct rim 8.1, and the others on the interiors of the grains with variable texture (2.2, 3.1, 5.1 and 6.2; Table 2). The latter four analyses of the 504±7 Ma population show a classic chondrite normalised REE pattern for zircons grown during eclogite facies metamorphism, with depressed HREE abundances (50–10 times chondrite) and devoid of any Eu anomaly (Fig. 4). For a rock like the gabbroic/doleritic protolith of sample 93222–0004, this pattern indicates zircon growth under eclogite facies, because plagioclase was absent (hence no negative Eu-anomaly) and there was abundant garnet present (resulting in the depressed HREE; Rubatto, 2002; Gilotti et al., 2004; Nutman et al., 2008). The pattern for rim analysis 8.1 does not display a negative Eu anomaly, but there is a steady increase in the HREE relative to reference chondrite, rather than the flat pattern seen for the other analyses. This suggests its growth in a slightly different environment (some garnet breakdown?) at the end of the eclogite facies metamorphic event.

3.4 Howqua River and Melville Point zircon U–Pb dating

Both the sample HQ9 from Howqua River and sample MVP from Melville Point yielded morphologically-similar zircon populations with the same age spectra, thus they are considered together here. CL images of representative zircons from the sample MVP are shown in Fig. 5. The grains all show abraded exteriors, thus they are devoid of in situ growth of metamorphic zircon. This is in keeping with the low grade (maximum greenschist facies) overprint on these rocks (Offler et al., 1998; Spaggiari et al., 2002b). The majority of the grains are oval or somewhat rounded prisms, whereas a minority of the grains in both samples are almost euhedral prisms. The rounded grains tend to show complex internal structures in CL images. Some have distinct cores and overgrowths, and truncation of zonation at the margins is common. Thus for example MVP grain 3 displays a Mesoproterozoic core, a broad Neoproterozoic rim, plus an outer thin partial rim, too narrow to be dated with a SHRIMP analytical spot (Fig. 5). For other rounded grains, zoning in old components have been truncated by abrasion at the grain boundary, such as for Neoarchean MVP grain 16 (Fig. 5). These grains give Precambrian ages, particularly latest Neoproterozoic (600–550 Ma), Mesoproterozoic (1100–1000 Ma), with a lesser number of older grains present (Table 1; Fig. 6). The rarer more euhedral, prismatic grains show oscillatory zoning parallel to the grain margins, but generally are dark in the CL images, due to higher U and Th content. These grains give the youngest ages (Table 1; Fig. 6), and will provide the best constraint on the maximum age of deposition of the Melville Point and Howqua River sandstones. Therefore SHRIMP analyses were focussed on these younger grains, with repeat analyses undertaken on several of them, in order to establish their age(s) with greater confidence. However, the commonly high U and Th content means that modern loss of radiogenic Pb in these grains can be a problem, giving apparent $^{206}Pb/^{238}U$ ages younger than the $^{207}Pb/^{206}Pb$ ages (with the latter being less precise at the early Phanerozoic). Thus high U (>500 ppm) grains and sites gave the youngest, scattered ages, which is regarded as due to later isotopic disturbance in grains with the most structural damage from radioactive decay. Grains with less U (300–100 ppm) yielded concordant ages at ~500 Ma. Assessments of the ages of these grains are as follows: for HQ9 505±34 (2σ), 499±17 (95% confidence), 496±32 (2σ), 495±19 (95% confidence), 495±34 (2σ) Ma and for MVP 510±10 (95% confidence), 484±15 (95% confidence) Ma. (2σ used when only 1 analysis was undertaken on a grain.) This suggests a maximum depositional age of lowermost Ordovician (Tremadocian) for the turbidite fan in the eastern Lachlan Fold Belt. The remaining signature of older grains in these sandstones is classic eastern Gondwanan, with a predominance of 600–550 and 1100–1000 Ma grains (Fig. 6).

4. Discussion

The main issue to be considered is the development of the Delamerian Fold Belt and its relationship to the deposition of a vast turbidite fan in the Lachlan Fold Belt. Earlier ideas on the Cambrian–Ordovician paleogeography of southeastern Australia regarded the Delamerian Orogeny as a major episode of shortening, which thickened crust of the Delamerian Fold Belt and thereby provided a repeatedly-recycled source for deposition of the widespread siliciclastic turbidite fan (Turner et al., 1996; Fergusson and Tye, 1999). Various plate tectonic scenarios have been given for the Delamerian Orogeny including an island arc – passive margin collision as exemplified by the geology of western Tasmania (Berry and Crawford, 1988; Meffre et al., 2000). Some recent work has raised questions about the vast Ordovician turbidite fan having formed adjacent to the Delamerian Fold Belt. It has been suggested that the deep-marine Ordovician siliciclastic rocks of the Lachlan Fold Belt formed one or more allochthonous terranes that have been transported along the East Gondwana margin (Baillie, 1985; Glen, 2005; Glen et al., 2009). Others suggested that long-distance sedimentary transport of sediment from the East – West Gondwana Pan-African collision zones thousands of kilometres away provided an explanation for the detrital U–Pb zircon ages in the Ordovician and other siliciclastic successions (Squire et al., 2006; Maidment et al., 2007; Williams and Pulford, 2008).

4.1 Delamerian(–Tyennan) Orogeny

From the space-time diagram for selected regions in southeastern Australia (Fig. 7) it is clear that the Delamerian (–Tyennan) Orogeny has substantial variations in terms of the timing and nature of events. Numerous intermittent episodes of rifting are associated with the formation of the Adelaide Rift Complex in South Australia dating from ~800 Ma onwards (Preiss, 2000). A 600–580 Ma passive margin has been recognised along the East Gondwana margin in the Delamerian Fold Belt (Direen and Crawford, 2003a, b). Widespread magnetic anomalies associated with volcanic successions with tholeiitic to alkaline magmatic affinities in the Koonenberry belt, western Victoria and in Tasmania have been interpreted as equivalent to the thick mafic volcanic units of seaward-dipping reflector sequences in volcanic passive margins.

A marked change in provenance has accompanied the Delamerian Orogeny as shown by the U–Pb detrital zircon ages published by Ireland et al. (1998). Prior to ~545 Ma, detrital zircon ages show that the Adelaide Rift Complex had various sources, such as a late Mesoproterozoic zircon source reflected in samples from the Bonney Sandstone and Marino Arkose, whereas other samples, such as the Niggly Gap beds and Mitcham Quartzite, were derived mainly from the adjoining Gawler craton. In contrast, a sample from the Kanmantoo Group shows a substantial 700–500 Ma age peak reflecting the input from a new source. Samples from the Normanville Group, the shallow marine succession underlying the Kanmantoo Group, reflect mainly the pre-545 Ma sources with derivation from the late Mesoproterozoic and 2000–1500 Ma peaks represented in the Heathersdale Shale sample and the 1830 Ma peak in the Mt Terrible Formation sample.

In Antarctica, development of a passive margin along the Ross Orogen has been placed at ~700 Ma with initiation of convergence along the margin occurring at ~580 Ma, as shown by the incoming detritus containing abundant detrital zircon and muscovite with ages of 580–500 Ma (Goodge et al., 2002, 2004a, b). The convergence is related to the inception of a continental margin magmatic arc reflecting conversion of the passive to an active margin with more complicated events in Northern Victoria Land related to accretion of an island arc represented by the Bowers Supergroup (Gibson and Wright, 1985). New radiometric ages from the Delamerian Fold Belt in South Australia indicate an earlier start to convergence at ~545 Ma rather than at 514 Ma as previously considered (Turner et al., 2009). These authors further point out that the Normanville and Kanmantoo groups did not form in an extensional basin setting as part of the former passive margin but were part of a synorogenic basinal setting like the Byrd Group in Antarctica.

In northwestern New South Wales, a convergent margin assemblage is recognised for the Cambrian Mount Wright Volcanics and associated shallow-marine successions formed in arc to forearc environments with a deep-marine turbidite succession to the northeast (Teltawongee Group) considered the equivalent subduction complex (Greenfield et al., 2011). U–Pb SHRIMP ages show that this assemblage lasted for 20 million years at ~525–505 Ma. It was succeeded by intense compressional deformation in the Delamerian Orogeny at 505–500 Ma in contrast to its longer time interval in South Australia. The arc-forearc assemblage developed over the former passive margin succession. One sample from the Nundora Formation in the western Teltawongee Group shows ages of detrital zircons with a prominent peak at 1178±17 Ma and no ages in the interval 600–500 Ma, whereas another sample from the Cooper Mine Range Formation shows a significant peak at 700–500 Ma (Greenfield et al., 2010, p. 346) indicating the varied provenance of units prior to the Delamerian Orogeny.

The post-Delamerian succession in the Koonenberry belt shows a similar detrital zircon age signature as found in the Kanmantoo Group with prominent age spectra in the range 600–500 Ma and 1300–1000 Ma as shown by samples from the Mutawintji Group and the Warratta inlier. However, in detail both these samples show prominent peaks in zircon ages at ~580 Ma consistent with derivation from the late Neoproterozoic rifted volcanic succession including the Mt Arrowsmith Volcanics (Fig. 8a) (Greenfield et al., 2010, p. 346; Greenfield et al., 2011). Additionally, one sample is from the Bilpa Conglomerate, which is poorly sorted with clasts up to 1 m across that are dominantly metasandstone, metapelite and mafic volcanic rocks indicative of local basement rock types affected by the Delamerian Orogeny (Pahl and Sirorska, 2004; Greenfield et al., 2010). The Bipla Conglomerate is part of a coarse-grained delta deposit formed in continental and shallow marine environments (Pahl and Sirorska, 2004). Thus we contest interpretations that these "Pacific-Gondwana" detrital zircon signatures, as found in the matrix of the Bilpa Conglomerate, are indicative of long-distance sedimentary transport (see below; Squire et al., 2006; Williams and Pulford, 2008).

In western Victoria, substantial cover by younger units has obscured the geological relationships, especially in the eastern part of the Delamerian Fold Belt. The Delamerian Orogeny in western Victoria is associated with development of abundant plutonic rocks in the Glenelg Zone (Crawford et al., 2003). Collision of an island arc boninitic-tholeiitic assemblage with a rifted margin has been postulated and was followed by "post-collisional" igneous activity, similar to events in Tasmania (VandenBerg et al., 2000; Crawford et al., 2003). It has been argued that the Glenelg Zone represents part of the passive margin succession that has been accreted into an accretionary wedge at an east-dipping subduction zone associated with the boninitic island arc, similar to relationships proposed for Northern Victoria Land (Gibson et al., 2011). These authors argued that following collision of the east-directed subduction zone with the passive margin, a new west-dipping subduction zone initiated east of the accreted island arc and caused the widespread igneous activity in the Glenelg Zone in addition to the "post-collisional" igneous activity of the Grampians–Stavely Zone. Ar/Ar ages from the Moorambool Metamorphic Complex in the western Stawell Zone indicate exhumation of garnet amphibolites at ~500 Ma from this west-dipping Delamerian subduction zone (Miller et al., 2005). Cayley (2011) has reinterpreted the tectonic development of the Delamerian Fold Belt in western Victoria as an active east-facing margin developed in the Early to Middle Cambrian with intensification of deformation in the Late Cambrian associated with collision of an outboard microcontinent, “Van Dieland”, that includes Tasmania and its inferred northern subsurface continuation, the Selwyn Block forming a deeply buried metamorphic basement to much of central Victoria (Cayley et al., 2002).

In Tasmania, the Tyennan Orogeny is attributed to an island arc – passive margin collision with emplacement of an island arc ophiolite allochthon of Cambrian boninitic-tholeiitic rocks over a Neoproterozoic passive margin (Berry and Crawford, 1988; Crawford and Berry, 1992). Metamorphic complexes, such as the Franklin Metamorphic Complex, were formed by subduction of a thin passive margin, lacking a young thick sedimentary cover, under the island arc as has occurred in Oman (Meffre et al., 2000). These authors also suggested that exhumation of these metamorphic rocks was aided by extensional unroofing associated with formation of the post-collisional Dundas Trough.

The location of Tasmania prior to and during the Delamerian(–Tyennan) Orogeny has also been widely debated in the literature. Paleomagnetic data from northwest Tasmania imply that in the Early Ordovician western Tasmania was in its approximate present position with respect to East Gondwana (Li et al., 1997). From monazite ages indicating metamorphism and granitoid intrusion in the interval 1290–920 Ma, Berry et al. (2008) inferred that the west Tasmania terrane, including the South Tasman Rise, was most likely connected to the Transantarctic Mountains prior to rifting at ~580 Ma. Arguing from relationships in western Victoria and Northern Victoria Land and including constraints from the Southern Ocean passive margins, Gibson et al. (2011) developed a comprehensive model for the Ross–Delamerian Orogeny. They suggested that a common east-dipping subduction zone existed with eastwards subduction of the East Gondwana passive margin under an island arc from north to south in western Victoria, Northern Victoria Land and Tasmania. Collision resulted in flipping of the subduction zone with development of west-dipping subduction and magmatic activity during the Middle to Late Cambrian. Tasmania has therefore been displaced from an original position near South Victoria Land along major sinistral strike-slip faults, such as the Avoca Fault and it offshore southern continuation. Somewhat less strike-slip displacement of the west Tasmania terrane and the Selwyn Block was argued by Cayley (2011). Just how any largescale sinistral strike-slip transport is accommodated in mainland southeastern Australia is problematic.

In summary, the Delamerian Orogeny of southeastern Australia reflects initiation of a continental margin convergent assemblage from 545 Ma but most widely developed in the interval 520–505 Ma with emplacement of island arc allochthons in the later part of this interval. The context of west Tasmania along the continental margin is poorly resolved but with a possibility that it may have collided with the Delamerian active margin at ~505 Ma prior to formation of a post-collisional volcanic succession developed in the Mt Read Volcanics of western Tasmania, the Mt Stavely Complex and equivalent units in western Victoria and the Licola Volcanics along the eastern margin of the Melbourne Zone in central Victoria (Cayley, 2011). Alternatively, others have argued for a more remote location for Tasmania further southwards along the Ross Orogen prior to terrane displacement that may have involved considerable strike-slip displacement to its Ordovician setting in East Gondwana (Berry et al., 2008; Gibson et al., 2011).

The Delamerian convergent assemblage is unusual in several ways. It is characterised by the development of low to locally high-grade metamorphic complexes associated with granitoids but lacks widespread granodioritic-tonalitic plutonic rocks more typical of an eroded continental margin arc. These are thought to occur along the western margin of the Ross Orogen but are presently covered by the East Antarctic ice sheet. Their presence has been inferred by abundant 600–500 Ma detrital zircons in siliciclastic rocks of the exposed part of the orogen (Goodge et al., 2002, 2004a, b; see below). For southeastern Australia, the connection between the Ross and Delamerian orogens is complicated by island arc collision in western Tasmania and suggested for western Victoria. We favour the idea of Cayley (2011) that the Delamerian Orogeny at 505–500 Ma was caused by collision of west Tasmania, and associated fragments, accompanied by development of new island arc systems in the paleo-Pacific Ocean such as the Macquarie Arc (Fig. 9).

4.2 Potential sources of the Ordovician turbidites?

Paleogeographic reconstructions assuming the present-day spatial arrangements between orogenic belts show that the source of the Ordovician turbidites was in the Delamerian Fold Belt to the west and south (Cas, 1983). This is consistent with available paleocurrent data indicating westward derivation for parts of the turbidite fan (Powell, 1984; Fergusson and VandenBerg, 2003). A source in the Delamerian Fold Belt is also consistent with sedimentary petrography of sandstones in the succession with abundant quartz, minor feldspar and minor metamorphic detritus (Fergusson and Tye, 1999). Sr isotopic data from whole rock samples of the Ordovician turbidites are also appropriate for sources such as the Kanmantoo Group in the Delamerian Fold Belt and igneous rocks in western Tasmania (Gray and Webb, 1995).

Detrital mica ages within the Ordovician sandstones have been determined in several Ar/Ar isotopic studies. Sandstone samples from the southern Bendigo Zone and the central part of the Tabberabbera Zone have detrital muscovite ages in the range 512–480 Ma (Turner et al., 1996). From sandstones in the Stawell and Bendigo zones, Foster et al. (1998) showed muscovite ages range 580–470 Ma with most ages in the interval 510–500 Ma. These data are in agreement with the timing of peak metamorphism in the Delamerian Fold Belt (Foden et al., 2006; Turner et al., 2009) and provide further support for a provenance connection between the Delamerian mountain chain and the Ordovician turbidites.

The source of the vast Ordovician turbidite fan of the Lachlan Fold Belt presents a dichotomy when the detrital zircon U–Pb ages are taken into account because of the lack of an obvious source of the 600–500 Ma and 1300–1000 Ma zircons. The samples from the Howqua River and Melville Point are considered informative given that these sandstones are earliest Ordovician and the least mature compositions recorded from the Ordovician turbidites, with higher feldspar and lithic contents (Fergusson and Tye, 1999). These samples show the common detrital zircon age spectra determined from other Ordovician sandstones with significant age intervals at 600–500 Ma and 1300–1000 Ma (Fig. 8b-g) (Ireland et al., 1998; Williams, 1998; Fergusson and Fanning, 2002). This signature has been widely recorded from sandstones in Australia. For example, two samples from the Triassic Hawkesbury Sandstone in the Sydney region are dominated by the 700–500 Ma detrital zircons, considered the "Pacific–Gondwana igneous component", and regarded as exotic and derived from East Antarctica (Sircombe, 1999). Alternatively, several authors have related these detrital zircon signatures to long-distance transport of significant volumes of sediment from the East African Orogen as discussed in the next section (Squire et al., 2006; Maidment et al., 2007; Williams and Pulford, 2008).

4.3 Is the East African Orogen the only possible source for the Ordovician turbidites?

It has been argued that the widespread detrital zircon age ranges of 1200–900 Ma and 650–550 Ma in Cambrian to Ordovician quartz-rich sandstones in different parts of Gondwana reflect erosion of the Transgondwanan Supermountain which was formed from the East African Orogen during collision of West and East Gondwana (Squire et al., 2006). Widespread deposition of mature quartz-rich sandstones of Cambrian to Ordovician ages in North Africa, the Middle East and displaced terranes in Europe have been related by Meinhold et al. (2013) to erosion and rejuvenated uplift in parts of the Transgondwanan Supermountain. It is also possible that the similar detrital ages reported by Cawood et al. (2007) in early Palaeozoic sandstones along the Tethyan margin of India are another component of this Gondwana super-fan derived from the Transgondwanan Supermountain. In the Ross Orogen and Delamerian and Lachlan fold belts similar signatures are common in quartz-rich sandstones and include the Ordovician turbidites which Squire et al. (2006) argued were part of the Gondwana super-fan derived by very long distance sedimentary transport from the southern part of the East African Orogen. As noted above a similar signature also occurs in Triassic quartz-rich sandstones of the Hawkesbury Sandstone in the Sydney Basin which indicates a markedly long-term uniformity of the apparent long-distance source. We argue that such long-distance transport suggested for the Ordovician turbidites is unlikely.

Does the Transgondwanan Supermountain (i.e. the East African Orogen) represent the only potential source of these Pan-African and Grenvillean detrital zircon signatures? Both these zircon age ranges are associated with supercontinent assembly. The Grenvillean and Pan-African zircon sources are widespread in the main continents as shown by Rino et al. (2008) and Voice et al. (2011). Although Antarctica lacks a significant peak in granitoid activity in the 1500–1000 Ma range (Condie et al., 2009), it has been argued from the prominence of the Grenvillean detrital zircons in Early Palaeozoic sedimentary rocks of the Ross Orogen (Goodge et al., 2002, 2004a) and also on the other side of the continent (Prince Charles Mountains) in Permian siltstone that the interior of East Antarctica must include a widespread and significant Grenville source(s) (Veevers and Saeed, 2008, 2011). Grenvillean sources are common in Africa and South America (Rino et al., 2008) and also running across Australia (Myers et al., 1996). The same logic also applies to the Pan-African detrital source which must have been even more widely developed throughout much of East Antarctica (Veevers and Saeed, 2008, 2011).

4.4 Derivation of the Ordovician turbidites from East Antarctica

Given that the East African Orogen cannot be considered a unique source for the 1300–1000 Ma and 600–500 Ma detrital signatures found in the Ordovician turbidites, we examine our data in comparison to data presented for other early Palaeozoic sandstones in southeastern Australia. In sample HQ9, two significant clusters occur within the age range 600–480 Ma, one with six grains with ages in the range 510–490 Ma (four of these grains have been checked by duplicate analyses). The other cluster is in the range 560–520 Ma with eight ages. In sample MVP, the youngest significant cluster at 510–480 Ma is based on only two grains (one has a duplicate analysis and the other triplicate analyses). Sample MVP also has a significant cluster in the range 560–520 Ma with 10 ages. Thus both samples show, even if only weakly in MVP, input from a Delamerian detrital zircon signature matching most likely igneous sources in the Delamerian Fold Belt. Cayley (2011) has used the post-collisional volcanic suites of western, central Victoria and Tasmania as a marker horizon (Fig. 9) and these rocks also provide a potential source of the youngest zircons in the Ordovician turbidites. Their relatively restricted distribution in the Ordovician turbidites shows that these volcanic successions were largely buried. Particularly pertinent, is the detrital ages of zircons in the proximal Bilpa Conglomerate of western New South Wales (Fig. 8a) that show most zircons derived within the Delamerian Fold Belt consistent with abundant locally derived metamorphic clasts (see above). A significant population of ~510–485 Ma zircons was noted by Squire et al. (2006) as reflecting a late Cambrian to earliest Ordovician zircon-bearing source in the Adaminaby Group which we would attribute to a Delamerian Fold Belt source.

The issue to be resolved is the source of the 600–520 Ma zircons, in particular the 560–520 Ma zircons found in HQ9 and MVP, as well as being widely distributed in other Ordovician and Late Cambrian sandstones (Fig. 8). A likely source is considered to be the now covered continental margin arc of the Ross Orogen of East Antarctica. The northernmost part of this occurs east of the Mertz Glacier in Terre Adelie Land (Fig. 1) and has been shown to include granitic rocks of 580–575 Ma and 515–500 Ma (Goodge and Fanning, 2011). Further south in the Central Trans-Antarctic Mountains, clastic rocks of the Douglas Conglomerate and Starshot Formation include abundant detrital zircons in the range 600–510 Ma as well as a significant Grenville component (Goodge et al., 2002, 2004a). It is clear that the Douglas Conglomerate represents a synorogenic conglomerate derived from the adjacent Ross Mountains and is analogous to the Bilpa Conglomerate of northwestern New South Wales. Additionally, Gibson et al. (2011) noted that another potential source for the 600–500 Ma zircons in the Kanmantoo Group was from erosion of the collided Cambrian boninitic island arc in western Victoria and Tasmania; in particular this would apply to the 560–520 Ma zircons in samples HQ9 and MVP. Similarly, detrital zircons in the range 600–570 Ma may have been derived from volcanic rocks associated with the former passive margin as in the Koonenberry belt, western Victoria and western Tasmania.

It is clear from Fig. 8 that different samples reflect a variation in the Delamerian component in the Ordovician turbidite sandstones. This is also apparent from a close inspection of the detrital zircon ages given by Squire et al. (2006). A sample from the Middle Cambrian Monegeetta Shale along the eastern margin of the Bendigo Zone, which is now included in the Knowsley East Formation (Crawford et al., 2003), is dominated by zircons in the age interval 550–500 Ma with over half the ages reported by Squire et al. (2006, Appendix A) in the range 520–480 Ma (Fig. 8h). Given than the Monegeetta Shale is a hemipelagic deposit overlying Cambrian boninitic-tholeiitic island arc basement it is more likely that these zircons were derived from an exposed part of these intraoceanic island arc units rather than the distant continent. In contrast, two samples from the St Arnaud Group, one sample from the Leviathan Formation and one sample from the Albion Formation all in the Stawell Zone (Fig. 8i-l) contain only one zircon in each sample with ages in the interval 520–490 Ma indicating that these sediments have only very minor Delamerian input. A common source for these samples is unlikely, as one of them, sample ST18, contains only two grains with ages in the interval 610–550 Ma and one grain with an age of 512 Ma indicating that even the Ross igneous source was minimal.

4.5 Paleogeographic reconstruction

We therefore favour a reconstruction of the earliest Ordovician paleogeography in which the Ross–Delamerian mountain range provides the main source of detrital zircons to the Ordovician turbidite fan (Fig. 9). Thus the Delamerian and Lachlan fold belts are considered to have developed adjacent to each other although this does not rule out limited strike-slip offsets and rotations of regions within the Lachlan Fold Belt during the complicated Silurian to Carboniferous tectonic development of that belt (e.g. Fergusson, 2010). Significantly, detrital zircon ages are interpreted as reflecting an East Gondwana provenance (<1000 km) rather than indicating long-distance fluvial transport (~5000 km) from the Transgondwanan Supermountain. Although many sandstones contain common elements to their detrital zircon age spectrums closer analysis of these shows that individual samples are varied and reflect different provenances even for samples from the same rock unit (e.g. samples ST21 and ST18 of Squire et al., 2006, from the St. Arnaud Group). Nevertheless, given the thick ice sheet covering the East Antarctic bedrock uncertainty remains about just how far sediment has travelled from potential sources in East Antarctica. Our preference is that the Ross Mountains favoured by Goodge et al. (2002, 2004a) is the most likely source.

The Ordovician turbidites form part of a major submarine fan estimated by Fergusson and Coney (1992) as having an area of 1.2 x 10^6 km^2 in the undeformed state and comparable to the area of the present-day Bengal fan at 3 x 10^6 km^2 and larger than other fans such as the Mississippi and Amazon fans at 3 x 10^5 km^2 (Bouma et al., 1985). Bengal Fan deposition was initiated in the Eocene from uplift of the Himalayas and Tibetan Plateau with sediment mainly transported along the Ganges and Brahmaputra rivers to a confluence at the head of the Bay of Bengal and onto the submarine fan (Curray et al., 2003). The turbidite fan prograded southwards as shown by a middle Eocene to upper Miocene hiatus between Bengal Fan sediments and the pre-fan sediments at DSDP site 215 and a longer hiatus at DSDP site 211 in contrast to fan deposition dating from near the time of collision in the Indoburman Ranges and the Andaman–Nicobar Ridge much closer to the source in the Bengal Basin (Curray et al., 2003, fig. 6). In contrast, the earliest Ordovician depositional ages for the turbidite fan in the Howqua River and Melville Point indicate rapid progradation of the turbidite fan, which following the interpretation of Fergusson (2009) formed two major lobes either side of the nascent Macquarie Arc (Fig. 9). This is consistent with the discovery of detrital zircons of similar ages to those of the Ordovician turbidites in a sample from the Early Ordovician Mitchell Formation near the base of the Macquarie arc succession (Glen et al., 2011). As for the Bengal Fan, considerable sediment was deposited prior to major fan progradation as shown by inferred late Cambrian turbidite deposition in the Stawell Zone of western Victoria. In southeastern Australia, major turbidite progradation followed collision of VanDieland with the Delamerian active margin with development of a paleogeography favouring diversion of vast quantities of sediment into one part of the adjacent paleo-Pacific Ocean (Fig. 9).

5. Conclusions

A CL-guided SHRIMP U–Pb zircon age of 504±7 Ma is consistent with earlier ages reported by Black et al. (1997) and Berry et al. (2007) for metamorphism in the Franklin Metamorphic Complex of western Tasmania. Based on the depressed HREE and lack of a Eu anomaly in the zircons this event is the eclogite facies metamorphism. These data are consistent with an island arc – passive margin collision at this time and from regional considerations was probably the collision of western Tasmania and associated fragments with the active Delamerian margin of East Gondwana as argued by Cayley (2011).

Final deformation in the Delamerian Orogeny accompanied formation of a widespread Ordovician turbidite fan comparable in undeformed extent to the present-day Bengal Fan. Sandstones from this deposit have distinctive 600–500 Ma and 1300–1000 Ma detrital zircon age intervals that match the times of Gondwanan and Rodinian supercontinent formation respectively. These sandstones have previously been considered part of the Gondwanan super-fan derived from the Transgondwanan Supermountain (i.e. the East African Orogen). However, evidence from detrital zircon analyses undertaken by Veevers and Saeed (2008, 2011) indicates that these detrital age intervals are potentially derived from much of the interior of East Antarctica. From a detailed comparison of detrital zircon spectra in Early Palaeozoic sandstones of southeastern Australia we have interpreted the provenance of the Ordovician turbidites as within East Gondwana most likely from the accreted island arc and magmatic arc associated with development of the Ross Orogen in East Antarctica.

Acknowledgements

We thank Ross Cayley and George Gibson and for helpful discussions. We also thank the Australian Academy of Sciences for Japan travel grant to undertake SHRIMP analyses at Hiroshima University. Hiroshima University provided generous access to analytical facilities. Geoscience Australia is thanked for permitting further studies on Franklin Metamorphic Complex sample 9322-0007. This work was also supported by the GeoQuEST Research Centre at the University of Wollongong. The manuscript has been significantly improved following the advice of the journal reviewers (Sebastien Meffre and an anonymous reviewer).

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Figure captions

Fig. 1. Gondwana reconstruction after de Witt et al. (1988) with modifications after Myers et al. (1996), Squire et al. (2006), Gray et al. (2008), Boger (2011) and Meinhold et al. (2013). Major cratonic units are highlighted in East Gondwana and with subdivisions of Australia and East Antarctica from Myers et al. (1996) and Boger (2011) respectively. Gondwana super-fan (GFS) and black arrows showing sediment movement pattern in north Africa from Squire et al. (2006). Black arrows shows sediment transport path required if Ordovician turbidites in the Lachlan Fold Belt of southeast Australia were derived from the East African Orogen (1) as suggested by Squire *et al.* (2006) and Williams and Pulford (2008) and shorter sediment transport route from East Antarctica (2). Path 3 provides an alternative source for early Palaeozoic sandstones on the Tethyan margin of India that may have been derived from the East African Orogen. The Lachlan Fold Belt in southeast Australia is divided into the Ordovician turbidite succession and the Ordovician Macquarie Arc. The Crohn-Mawson cratons are an amalgam of Precambrian and early Palaeozoic units largely concealed beneath the East Antarctic ice sheet; alternative representations of these units are given by Boger (2011) and Veevers and Saeed (2011). Abbreviations: AFMB–Albany-Fraser-Musgrave belt, DFB–Delamerian Fold Belt, GSF–Gondwanan super-fan, GSM–Gamburtsev Subglacial Mountains, NAC–North Australian Craton, RO–Ros Orogen, RP–Río de la Plata Craton, SAC–South Australian Craton, SF–São Francisco Craton, TC–Tanzania Craton, WAC–West Australian Craton.

Fig. 2. (a) Southeast Australia with Early Palaeozoic units of the Lachlan Fold Belt and the extent of Delamerian Fold Belt (also known as the west Tasmania terrane in western Tasmania). Locations of the Franklin Metamorphic Complex, Melville Point and Howqua River are indicated. The northeast Tasmania terrane consists of an Ordovician to Devonian succession with abundant granite and is considered equivalent to the eastern Lachlan Fold Belt (Cayley, 2011). Compiled and interpreted from data supplied by Geoscience Australia. Subsurface extensions of the Avoca and Moyston faults and the Hay–Booligal Zone inferred from magnetic data (after Hallet et al., 2005). (b) Location map for sample HQ9 (GDA94, 444685 5883685) in the Howqua River (Dev volcs = Devonian volcanic rocks), see (a) for location. (c) Location map for sample MVP (GDA94, 246515 6031165) at Melville Point, see (a) for location.

Fig. 3. CL images of zircons from the Franklin Metamorphic Complex eclogite sample.

Fig. 4. Franklin Metamorphic complex zircon data. (a) $^{238}U/^{206}Pb$ – $^{207}Pb/^{206}Pb$ concordia diagram (errors are portrayed at the 2σ level). (b) Chondrite-normalised REE plot. Field of common igneous zircon after Belousova et al. (2002) shown in gray.

Fig. 5. CL images of representative zircons from sandstone samples MVP (Melville Point).

Fig. 6. Sandstone sample detrital zircon ages; (a) Howqua River and (b) Melville Point, with the main figures being $^{238}U/^{206}Pb$ – $^{207}Pb/^{206}Pb$ concordia diagrams (errors are portrayed at the 2σ level) and the insets being $^{206}Pb/^{238}U$ age histograms and cumulative frequency distributions for the youngest grains. (c) $^{206}Pb/^{238}U$ age histograms and cumulative frequency distributions for the grains with close to concordant ages from both samples.

Fig. 7. Time-space plot showing the Delamerian Fold Belt and selected zones of the Lachlan Fold Belt in the Neoproterozoic to Late Ordovician. Data compiled from the following sources: Delamerian Fold Belt in South Australia – Preiss (2000), Foden et al. (2006), Turner et al. (2009); Grampians–Stavely Zone – Crawford et al. (2003), Cayley (2011); western Tasmania – Black et al. (1997), Turner et al. (1998); Koonenberry belt – Greenfield et al. (2010, 2011); western Stawell Zone – Miller et al. (2005), Squire and Wilson (2005); Tabberabbera Zone – Fergusson and VandenBerg (2003); Tomakin – Glen et al. (2004), Prendergast (2007).

Fig. 8. Comparison of detrital zircon age plots of various Late Cambrian units and Ordovician turbidites from the Lachlan Fold Belt, via $^{206}Pb/^{238}U$ age histograms and cumulative frequency distributions. (a) Matrix from the Bilpa Conglomerate of the Mutawintji Group, Koonenberry belt, northwestern New South Wales (Greenfield et al., 2010, fig. 127b, p. 346). (b) Melville Point sample (this paper). (c) Howqua River sample (this paper). (d) Ordovician Bumballa Formation (turbidites), Shoalhaven River, Lachlan Fold Belt in eastern New South Wales (Fergusson and Fanning, 2002, sample 514U1). This shows a prominent Delamerian signature at ~500 Ma and in the interval 570–520 Ma. (e) Ordovician turbidites from the upper Adaminaby Group, Genoa River, Lachlan Fold Belt in eastern Victoria (Fergusson and Fanning, 2002, sample CF256). Solid 600–500 Ma interval. (f) Quartzite from Ordovician turbidites of the Girilambone Group, Lachlan Fold Belt, central New South Wales (Fergusson et al., 2005, sample TH1). Mainly in range 600–520 Ma. (g) Quartzose sandstone from Ordovician turbidites of the Girilambone Group, Lachlan Fold Belt, central New South Wales (Fergusson et al., 2005, sample GG2). Delamerian – 550–500 Ma. (h) Middle Cambrian Monegeetta Shale, eastern margin of the Bendigo Zone, Lachlan Fold belt of central Victoria (Squire et al., 2006, sample ST46). 550–500 Ma – age of arc. (i) Cambrian Leviathan Formation, western Stawell Zone, Lachlan Fold Belt of western Victoria (Squire et al., 2006, ST23). 600–560 Ma interval. (j) Middle Cambrian Albion Formation, western Stawell Zone, Lachlan Fold Belt of western Victoria (Squire et al., 2006, ST43). Small grouping 600–550 Ma. (k) Late Cambrian St. Arnaud Group, Stawell Zone, Lachlan Fold belt of western Victoria (Squire et al., 2006, sample ST18). 3 grains in

range 600–500 Ma. (l) Late Cambrian St. Arnaud Group, Stawell Zone, Lachlan Fold belt of western Victoria (Squire et al., 2006, ST21). 11 in range 600–550 Ma.

Fig. 9. A speculative reconstruction of the East Gondwana margin at ~480 Ma in the earliest Ordovician showing southeast Australia and adjacent Antarctica. The “VanDieland” block includes Tasmania and adjacent submarine rises and the Selwyn Bock of central Victoria (after Cayley, 2011, fig. 6a). VanDieland has collided with the northern extension of the active margin of Northern Victoria Land (NVL). Further south a transform fault connects with the Leap Year Fault (LYF), the site of a recently active subduction zone. Reconstruction of the Ordovician turbidite fan of the Lachlan Fold Belt is with two arms of the fan shown on both sides of the Macquarie Arc and takes into account the inferred rotation of the arc around the inferred pivot point (i.e. Euler pole) and significant latest Ordovician to Carboniferous shortening (modified after Fergusson, 2009, fig. 3a). Approximate locations of samples HQ9 and MVP are given. The western margin of the recently active magmatic arc of the Ross Orogen provides a substantial input of detritus (arrows) into the Ordovician megafan in addition to the Delamerian Fold Belt. The post-collisional volcanic suite forms a potential marker traceable from western Victoria (Stavely Volcanic Complex) into central Victoria (Jamieson Volcanics), and Tasmania (Mt Read Volcanics) and shown as a dashed gray line (after Cayley, 2011). MF–Moyston Fault.

Tables

Table 1. Zircon U–Th–Pb data for the Tasmanian eclogite sample 9322-0007 (Franklin Metamorphic Complex) and Ordovician sandstone samples (MVP, Melville Point; HQ9, Howqua River).

Labels	site	U/ppm	Th/ppm	Th/U	f206%	$^{238}U/^{206}Pb$			$^{207}Pb/^{206}Pb$			age			method	conc%
9322-0007, Mt Franklin metamorphics																
1.1	r,h,ov,fr	76	0.8	0.010	0.77	12.126	±	0.537	0.0660	±	0.0033	507	±	22	U-Pb	
1.2	c,het,ov,fr	319	1.2	0.004	<0.01	12.196	±	0.345	0.0581	±	0.0021	509	±	14	U-Pb	
2.1	r,h,ov	414	1.4	0.003	0.16	12.306	±	0.274	0.0595	±	0.0013	503	±	11	U-Pb	
2.2	m,h,ov	173	3.0	0.018	<0.01	11.963	±	0.491	0.0585	±	0.0033	518	±	21	U-Pb	
3.1	e,h,ov	134	0.7	0.005	0.27	12.461	±	0.436	0.0626	±	0.0022	496	±	17	U-Pb	
4.1	m,sz,eq	206	6.4	0.031	0.21	12.314	±	0.314	0.0642	±	0.0017	500	±	12	U-Pb	
5.1	m,sz,eq	232	2.7	0.012	0.26	12.147	±	0.283	0.0618	±	0.0021	508	±	11	U-Pb	
5.2	e,h,ov	121	1.1	0.009	0.45	12.656	±	0.325	0.0629	±	0.0036	488	±	12	U-Pb	
6.1	e,sz,ov	376	1.5	0.004	0.23	12.182	±	0.243	0.0597	±	0.0019	508	±	10	U-Pb	
6.2	m,h,ov	122	0.7	0.006	0.36	12.568	±	0.301	0.0636	±	0.0029	491	±	11	U-Pb	
7.1	m,h,ov	226	1.4	0.006	0.19	12.172	±	0.336	0.0596	±	0.0016	509	±	14	U-Pb	
8.1	r,h,ov	65	0.3	0.004	0.82	13.218	±	1.030	0.0769	±	0.0066	460	±	35	U-Pb	
8.2	m,h,ov	245	0.7	0.003	0.31	12.096	±	0.344	0.0626	±	0.0025	510	±	14	U-Pb	
9.1	m,h,ov	202	2.0	0.010	0.36	12.094	±	0.310	0.0612	±	0.0029	511	±	13	U-Pb	
HQ9, Howqua River quartzite																
1.1	e,hd,eq	708	294	0.415	8.08	11.757	±	0.280	0.0572	±	0.0041	526	±	12	U-Pb	
2.1	c,h,ov	86	51	0.591	1.95	10.569	±	0.310	0.0479	±	0.0082	583	±	16	U-Pb	
3.1	e,osc,p/eq	162	118	0.726	0.21	10.977	±	0.373	0.0588	±	0.0024	562	±	18	U-Pb	
4.1	m,osc/h,ov	77	35	0.459	<0.01	5.817	±	0.182	0.0809	±	0.0060	1219	±	153	Pb-Pb	84
5.1	c,osc/hd,p	385	255	0.664	0.22	5.622	±	0.092	0.0714	±	0.0012	1055	±	16	U-Pb	
7.1	c,osc,p,fr	298	160	0.537	0.11	5.782	±	0.174	0.0743	±	0.0017	1050	±	47	Pb-Pb	98
8.1	c/m,osc,p,fr	202	48	0.237	0.45	5.558	±	0.206	0.0710	±	0.0019	1067	±	37	U-Pb	
9.1	m,osc/h,fr	190	136	0.714	<0.01	9.351	±	0.303	0.0630	±	0.0030	655	±	20	U-Pb	
10.1	m,osc,p,fr	159	103	0.647	0.64	12.505	±	0.427	0.0516	±	0.0052	496	±	16	U-Pb	
11.1	m,hd/osc,p	443	440	0.992	0.03	1.978	±	0.072	0.1856	±	0.0009	2703	±	8	Pb-Pb	98
15.1	e,osc,p,fr	169	151	0.894	0.26	3.203	±	0.093	0.1085	±	0.0026	1775	±	44	Pb-Pb	99
16.1	c/m,osc/hd,p	287	149	0.520	0.41	9.437	±	0.313	0.0587	±	0.0024	649	±	21	U-Pb	
17.1	c,osc,p	131	94	0.713	0.18	3.420	±	0.443	0.1069	±	0.0160	1747	±	303	Pb-Pb	95
18.1	e/r,osc,eq	93	54	0.583	0.02	5.766	±	0.133	0.0718	±	0.0029	1031	±	22	U-Pb	
19.1	e,osc,p,fr	162	30	0.185	0.07	5.426	±	0.357	0.0715	±	0.0048	1090	±	66	U-Pb	
20.1	m,hd/osc,p,fr	581	181	0.311	0.06	2.331	±	0.079	0.1459	±	0.0009	2298	±	10	Pb-Pb	100
21.1	e,osc/hd,p/fr	375	185	0.494	0.31	12.477	±	0.272	0.0551	±	0.0021	497	±	10	Pb-Pb	
21.2	m,osc/hd,p/fr	358	176	0.492	<0.01	12.274	±	0.413	0.0575	±	0.0021	505	±	16	Pb-Pb	
23.1	c,osc,ov	314	96	0.307	<0.01	11.536	±	0.247	0.0597	±	0.0044	536	±	11	U-Pb	
24.1	c,osc,ov	155	74	0.474	0.32	11.549	±	0.480	0.0573	±	0.0027	535	±	21	U-Pb	
25.1	e,osc,p,fr	202	105	0.520	0.60	11.401	±	0.341	0.0555	±	0.0046	542	±	16	U-Pb	
26.1	c,osc,p	169	65	0.387	0.07	5.545	±	0.243	0.0578	±	0.0034	1069	±	43	U-Pb	
27.1	m/c,osc,p,fr	172	145	0.843	4.66	7.006	±	0.276	0.0689	±	0.0087	860	±	32	U-Pb	
28.1	m,osc,p/r	729	467	0.640	7.15	6.524	±	0.254	0.0724	±	0.0097	997	±	300	Pb-Pb	92
29.1	c/m,osc,p	127	113	0.897	<0.01	5.532	±	0.195	0.0753	±	0.0052	1077	±	145	Pb-Pb	100
30.1	c,osc,p/ov	107	30	0.284	0.26	5.120	±	0.800	0.0477	±	0.0050	1150	±	167	U-Pb	
31.1	e,sz,ov	89	89	1.008	0.50	10.680	±	0.314	0.0549	±	0.0061	577	±	16	U-Pb	
32.1	e,p,osc,fr	270	141	0.522	0.15	11.563	±	0.348	0.0573	±	0.0015	535	±	15	U-Pb	
33.1	c,osc,p,fr	265	136	0.515	0.14	5.703	±	0.165	0.0733	±	0.0014	1042	±	28	U-Pb	
34.1	e,osc,p,fr	97	32	0.324	<0.01	5.655	±	0.177	0.0738	±	0.0068	1050	±	30	U-Pb	
35.1	c,osc,p,fr	120	42	0.347	<0.01	5.791	±	0.203	0.0758	±	0.0025	1089	±	68	Pb-Pb	94
36.1	c/m,osc,p,fr	160	92	0.577	0.16	6.154	±	0.168	0.0702	±	0.0026	971	±	25	U-Pb	
37.1	m,osc/hd,p,fr	307	105	0.343	<0.01	12.628	±	0.347	0.0566	±	0.0023	491	±	13	Pb-Pb	
37.2	e,osc/hd,p,fr	466	139	0.297	<0.01	12.421	±	0.378	0.0570	±	0.0035	499	±	15	Pb-Pb	
38.1	c,osc,p	541	339	0.627	0.05	11.963	±	1.110	0.0574	±	0.0018	518	±	46	U-Pb	
39.1	e,osc/h,fr	209	76	0.365	0.46	5.724	±	0.135	0.0709	±	0.0023	1038	±	23	U-Pb	
40.1	e,osc,fr	217	125	0.579	0.49	10.841	±	0.233	0.0544	±	0.0031	569	±	12	U-Pb	
41.1	e,osc,fr	139	42	0.299	0.04	5.285	±	0.145	0.0778	±	0.0016	1142	±	42	Pb-Pb	98
42.1	e,osc,p	481	285	0.593	0.18	12.044	±	0.248	0.0583	±	0.0018	514	±	10	U-Pb	
43.1	m,osc,p	292	306	1.051	0.24	12.263	±	0.436	0.0558	±	0.0019	505	±	17	U-Pb	
44.1	e,osc,fr	294	48	0.163	0.09	10.953	±	0.226	0.0579	±	0.0018	563	±	11	U-Pb	
45.1	e,osc,p,fr	147	64	0.438	0.56	11.467	±	0.456	0.0535	±	0.0052	539	±	21	U-Pb	
46.1	m,hd,eq	428	92	0.215	0.10	11.282	±	0.324	0.0589	±	0.0024	547	±	15	U-Pb	
47.1	m/c,osc,p	462	339	0.734	0.08	2.191	±	0.060	0.1854	±	0.0012	2702	±	11	Pb-Pb	90
48.1	c,osc,ov	408	218	0.534	0.66	9.466	±	0.369	0.0579	±	0.0041	647	±	24	U-Pb	

49.1	e,osc/hd,p	1237	981	0.793	7.70	12.541 ± 0.438	0.0637 ± 0.0159	495 ± 17	Pb-Pb	
49.2	e,osc/hd,p	831	468	0.563	0.14	11.558 ± 0.273	0.0569 ± 0.0009	535 ± 12	Pb-Pb	
50.1	e,osc,p	797	507	0.636	0.07	11.206 ± 0.227	0.0582 ± 0.0009	551 ± 11	U-Pb	
MVP, Melville Point sandstone										
1.1	e,osc,p,fr	157	113	0.720	0.12	3.478 ± 0.062	0.0991 ± 0.0016	1607 ± 30	Pb-Pb	101
2.1	m,osc,ov	228	97	0.427	0.04	2.967 ± 0.044	0.1150 ± 0.0009	1879 ± 15	Pb-Pb	100
3.1	c,osc,ov	304	224	0.736	<0.01	5.555 ± 0.272	0.0771 ± 0.0023	1122 ± 60	Pb-Pb	95
3.2	r,h,ov	97	104	1.071	0.33	10.971 ± 0.246	0.0588 ± 0.0034	562 ± 12	U-Pb	
3.3	r,h,ov	73	83	1.136	<0.01	11.109 ± 0.379	0.0640 ± 0.0024	556 ± 18	U-Pb	
4.1	e,osc,p,fr	208	111	0.535	0.26	5.695 ± 0.092	0.0725 ± 0.0013	1043 ± 16	U-Pb	
5.1	e/r,h,ov	232	2	0.008	<0.01	11.618 ± 0.195	0.0596 ± 0.0013	532 ± 9	U-Pb	
5.2	c,osc,ov	264	117	0.441	0.18	6.477 ± 0.143	0.0635 ± 0.0016	926 ± 19	U-Pb	
6.1	e,osc,p,tw	329	246	0.749	0.30	12.211 ± 0.188	0.0566 ± 0.0015	507 ± 8	U-Pb	
6.2	e,osc,p,tw	266	184	0.691	0.11	12.255 ± 0.275	0.0588 ± 0.0011	506 ± 11	U-Pb	
6.3	m,osc,p,tw	144	94	0.656	<0.01	11.915 ± 0.239	0.0619 ± 0.0014	520 ± 10	U-Pb	
7.1	e,osc,p	300	148	0.492	0.21	5.885 ± 0.186	0.0741 ± 0.0014	1043.7 ± 39.68	Pb-Pb	97
8.1	e,osc,p	290	164	0.564	0.12	11.465 ± 0.483	0.0576 ± 0.0021	539 ± 22	U-Pb	
8.2	m,osc,p	454	288	0.633	0.17	11.132 ± 0.426	0.0578 ± 0.0017	555 ± 20	U-Pb	
9.1	e,osc,ov	416	61	0.148	0.13	5.780 ± 0.091	0.0729 ± 0.0006	1029 ± 15	U-Pb	
10.1	e,osc/hd,p	1489	1146	0.770	5.41	16.349 ± 0.728	0.0569 ± 0.0030	383 ± 17	U-Pb*	
10.2	e,osc/hd,p	1203	1036	0.861	4.31	16.950 ± 0.502	0.0610 ± 0.0039	370 ± 11	U-Pb*	
11.1	e,osc/hd,p	870	195	0.224	0.51	6.985 ± 0.104	0.0723 ± 0.0011	862 ± 12	U-Pb	
12.1	m,osc,p	735	363	0.494	0.88	6.675 ± 0.125	0.0722 ± 0.0014	1044 ± 40	Pb-Pb	91
13.1	m,osc,p	254	229	0.899	0.06	9.129 ± 0.199	0.0595 ± 0.0013	670 ± 14	U-Pb	
14.1	e,osc,p	340	132	0.389	0.05	6.033 ± 0.159	0.0738 ± 0.0016	1035 ± 45	Pb-Pb	96
15.1	m,osc,ov	285	124	0.435	0.16	11.630 ± 0.257	0.0573 ± 0.0018	532 ± 11	U-Pb	
15.2	m,osc,ov	269	120	0.447	0.13	11.767 ± 0.188	0.0577 ± 0.0012	526 ± 8	U-Pb	
16.1	m,osc,ov	300	359	1.195	0.03	1.951 ± 0.162	0.1941 ± 0.0024	2777 ± 21	Pb-Pb	96
17.1	m,osc,ov	294	103	0.351	0.18	10.353 ± 0.274	0.0570 ± 0.0015	594 ± 15	U-Pb	
18.1	r,h,ov	411	12	0.030	0.37	10.463 ± 0.186	0.0577 ± 0.0014	588 ± 10	U-Pb	
19.1	m,osc,ov	474	72	0.152	0.15	9.843 ± 0.180	0.0650 ± 0.0016	624 ± 11	U-Pb	
20.1	m,osc,ov	158	134	0.851	<0.01	2.141 ± 0.095	0.1746 ± 0.0011	2602 ± 11	Pb-Pb	95
21.1	c,sz,ov	137	84	0.615	0.18	3.440 ± 0.064	0.1023 ± 0.0012	1666 ± 23	Pb-Pb	99
22.1	m,osc,p	309	146	0.472	0.18	12.933 ± 0.261	0.0550 ± 0.0014	480 ± 9	U-Pb	
22.2	m,osc,p	294	132	0.450	0.16	12.612 ± 0.387	0.0575 ± 0.0019	492 ± 15	U-Pb	
23.1	m,osc,p,fr	143	59	0.413	0.64	11.643 ± 0.422	0.0538 ± 0.0024	531 ± 19	U-Pb	
24.1	r,hd,p	947	1031	1.089	1.81	16.936 ± 0.303	0.0554 ± 0.0016	370 ± 6	U-Pb*	
25.1	c,osc,ov	263	73	0.280	0.38	6.319 ± 0.087	0.0640 ± 0.0028	947 ± 12	U-Pb	
26.1	m,h,ov	107	35	0.326	0.88	11.526 ± 0.281	0.0556 ± 0.0034	536 ± 13	U-Pb	
27.1	e,osc,ov	70	45	0.646	0.45	5.285 ± 0.108	0.0770 ± 0.0038	1120 ± 103	Pb-Pb	100
28.1	r,h,ov	132	56	0.426	0.29	11.884 ± 0.384	0.0604 ± 0.0018	521 ± 16	U-Pb	
29.1	e,osc,ov	132	74	0.565	<0.01	5.492 ± 0.111	0.0772 ± 0.0011	1127 ± 29	Pb-Pb	96
30.1	m,osc,ov	184	119	0.646	0.32	5.191 ± 0.113	0.0803 ± 0.0013	1205 ± 32	Pb-Pb	94
31.1	419±60	240	140	0.583	<0.01	14.877 ± 2.191	0.0654 ± 0.0108	419 ± 60	U-Pb*	

all uncertainties in the Table are given at 1 sigma.

Site: x.y, x=grain number, y=analysis number.

Grain and site character: p=prism, eq=equant/stubby, ov=oval, tw=twinned grain, fr=grain fragment

e=end analysis site, m=middle analysis site, rim=rim analysis site, c=core analysis site

CL imagery: osc=oscillatory finescale zoning, sz=sector zoning, het=heterogeneous, hd=non-luminescent

method: U-Pb=206Pb/238U age used, Pb-Pb=207Pb/206Pb age used, U-Pb*=young apparent age due to recent loss of radiogenic Pb

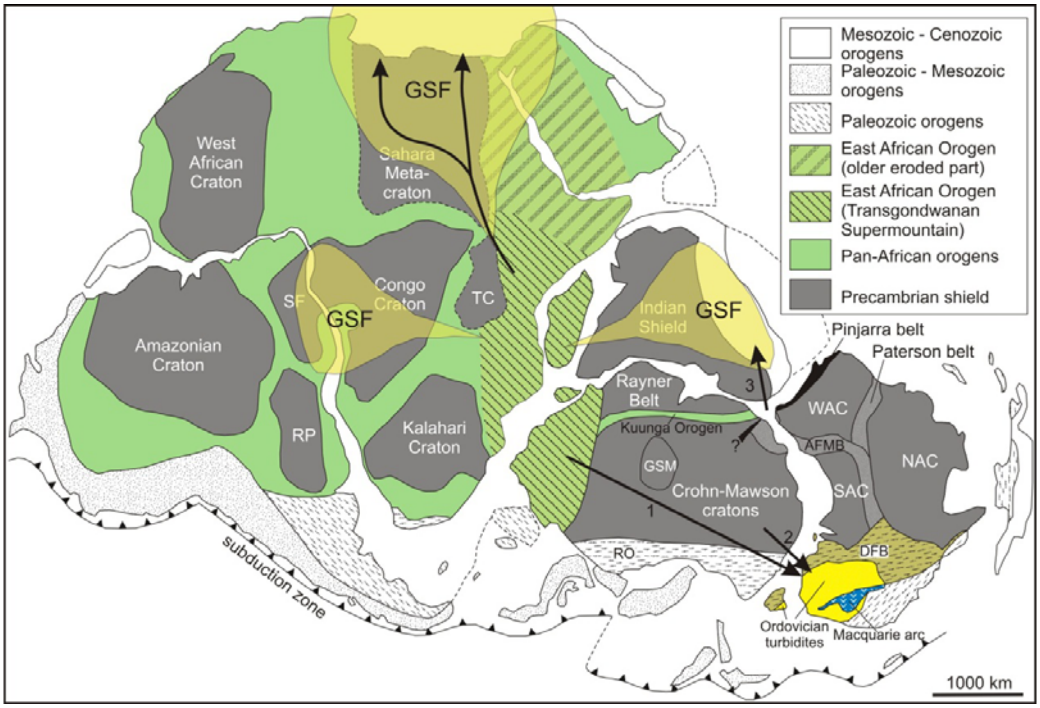
Common Pb correction: comm 206%= percentage of Pb that is non-radiogenic, based on measured 204Pb

and common Pb modelled as Cumming and Richards (1975) for likely age of rock

Table 2 REE analyses of the zircons from the eclogite sample, Franklin Metamorphic Complex.

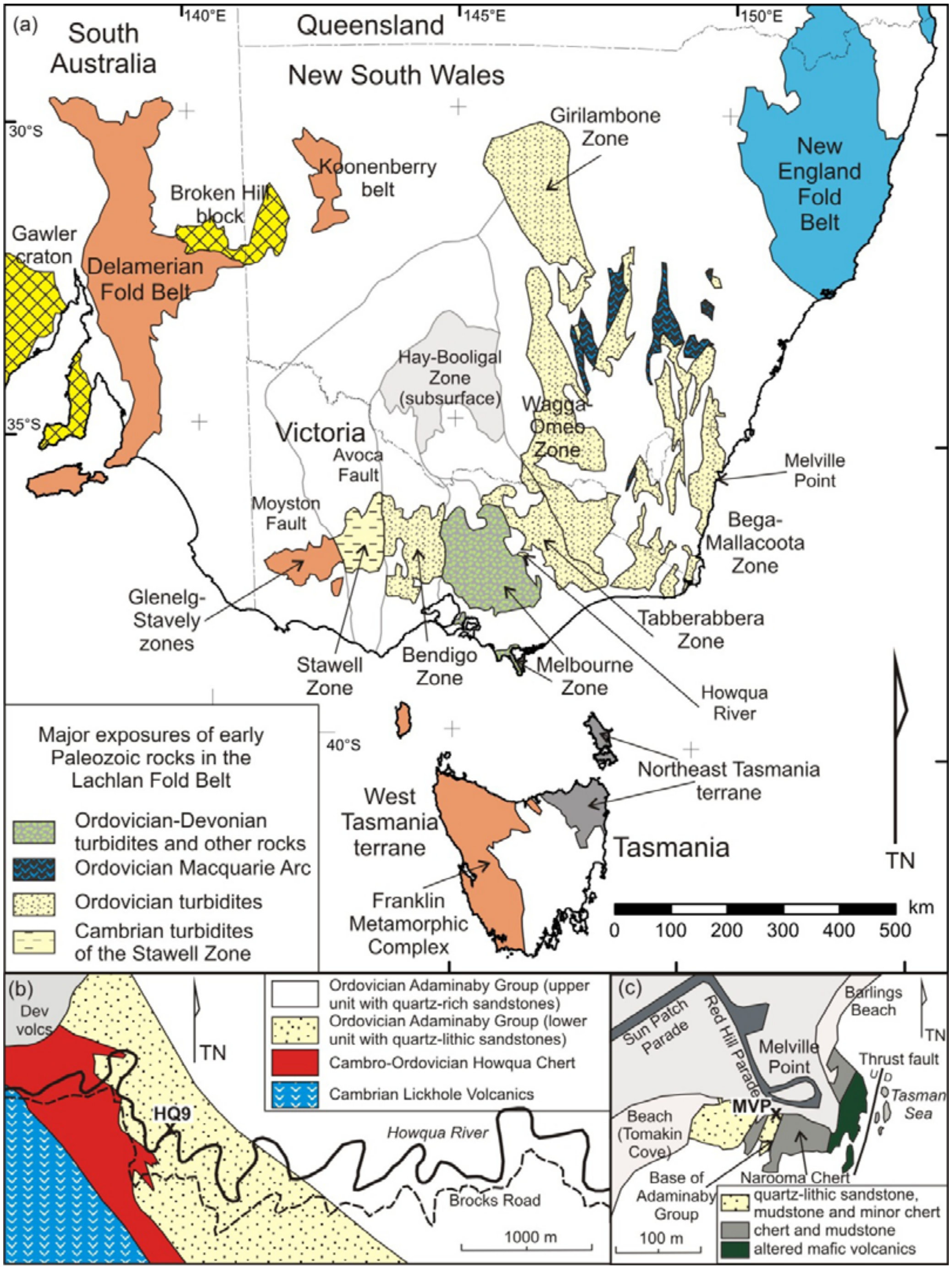
analysis	site	age	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu
9322-0007, Mt Franklin metamorphics																
2.2	m,h,ov	518	0.003	0.377	0.019	0.098	0.544	0.509	5.148	1.557	10.016	1.569	3.319	0.486	3.499	0.538
3.1	e,h,ov	496	0.003	0.244	0.014	0.072	0.439	0.398	3.670	0.956	5.786	1.015	2.636	0.427	3.353	0.544
5.1	m,sz,eq	508	0.051	0.521	0.029	0.163	0.463	0.419	3.372	0.913	5.783	1.051	2.667	0.411	2.881	0.431
6.2	m,h,ov	491	0.006	0.330	0.017	0.137	0.556	0.506	3.742	0.855	5.312	1.102	3.138	0.563	4.365	0.684
8.1	r,h,ov	460	0.232	1.330	0.053	0.896	0.603	0.401	2.295	0.625	5.147	1.541	5.985	1.295	10.9	2.100

grain morphology: eq=equant, ov=oval, fr=grain fragment, e=grain end, m=grain middle, r=overgrowth
CL imagery: sz=sector zoned, h=homogeneous
age is SHRIMP U-Pb age



Mesozoic - Cenozoic orogens
Paleozoic - Mesozoic orogens
Paleozoic orogens
East African Orogen (older eroded part)
East African Orogen (Transgondwanan Supermountain)
Pan-African orogens
Precambrian shield
GSF
Sahara Meta-craton
West African Craton
Amazonian Craton
SF
Congo Craton
TC
RP
Kalahari Craton
Indian Shield
Rayner Belt
Kuunga Orogen
GSM
Crohn-Mawson cratons
Pinjarra belt
Paterson belt
WAC
AFMB
NAC
SAC
DFB
RO
Ordovician turbidites
Macquarie arc
subduction zone
1000 km

Fig. 1.



(a)
South Australia
Queensland
New South Wales
140°E
145°E
150°E
30°S
35°S
40°S
Girilambone Zone
Koonenberry belt
Broken Hill block
Gawler craton
Delamerian Fold Belt
New England Fold Belt
Hay-Booligal Zone (subsurface)
Wagga-Omeo Zone
Victoria
Avoca Fault
Moyston Fault
Melville Point
Bega-Mallacoota Zone
Glenelg-Stavely zones
Stawell Zone
Bendigo Zone
Melbourne Zone
Tabberabbera Zone
Howqua River
Northeast Tasmania terrane
West Tasmania terrane
Tasmania
Franklin Metamorphic Complex
TN
km
0 100 200 300 400 500
Major exposures of early Paleozoic rocks in the Lachlan Fold Belt
Ordovician-Devonian turbidites and other rocks
Ordovician Macquarie Arc
Ordovician turbidites
Cambrian turbidites of the Stawell Zone
(b)
Dev volcs
TN
HQ9
Howqua River
Brocks Road
1000 m
Ordovician Adaminaby Group (upper unit with quartz-rich sandstones)
Ordovician Adaminaby Group (lower unit with quartz-lithic sandstones)
Cambro-Ordovician Howqua Chert
Cambrian Lickhole Volcanics
(c)
Sun Patch Parade
Red Hill Parade
Melville Point
Barlings Beach
TN
Thrust fault
U D
Tasman Sea
MVP
Beach (Tomakin Cove)
Base of Adaminaby Group
Narooma Chert
100 m
quartz-lithic sandstone, mudstone and minor chert
chert and mudstone
altered mafic volcanics

Fig. 2.

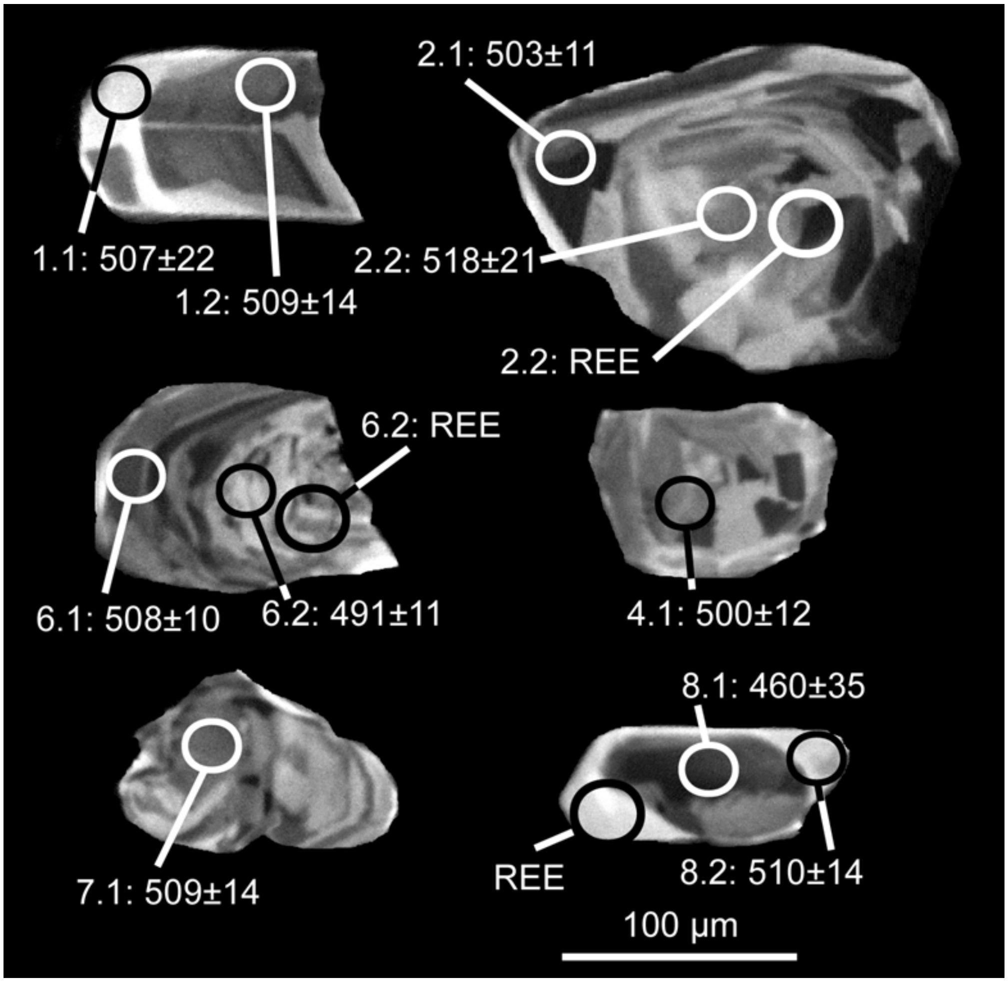

2.1: 503±11
1.1: 507±22
1.2: 509±14
2.2: 518±21
2.2: REE
6.2: REE
6.1: 508±10
6.2: 491±11
4.1: 500±12
8.1: 460±35
7.1: 509±14
REE
8.2: 510±14
100 μm

Fig. 3.

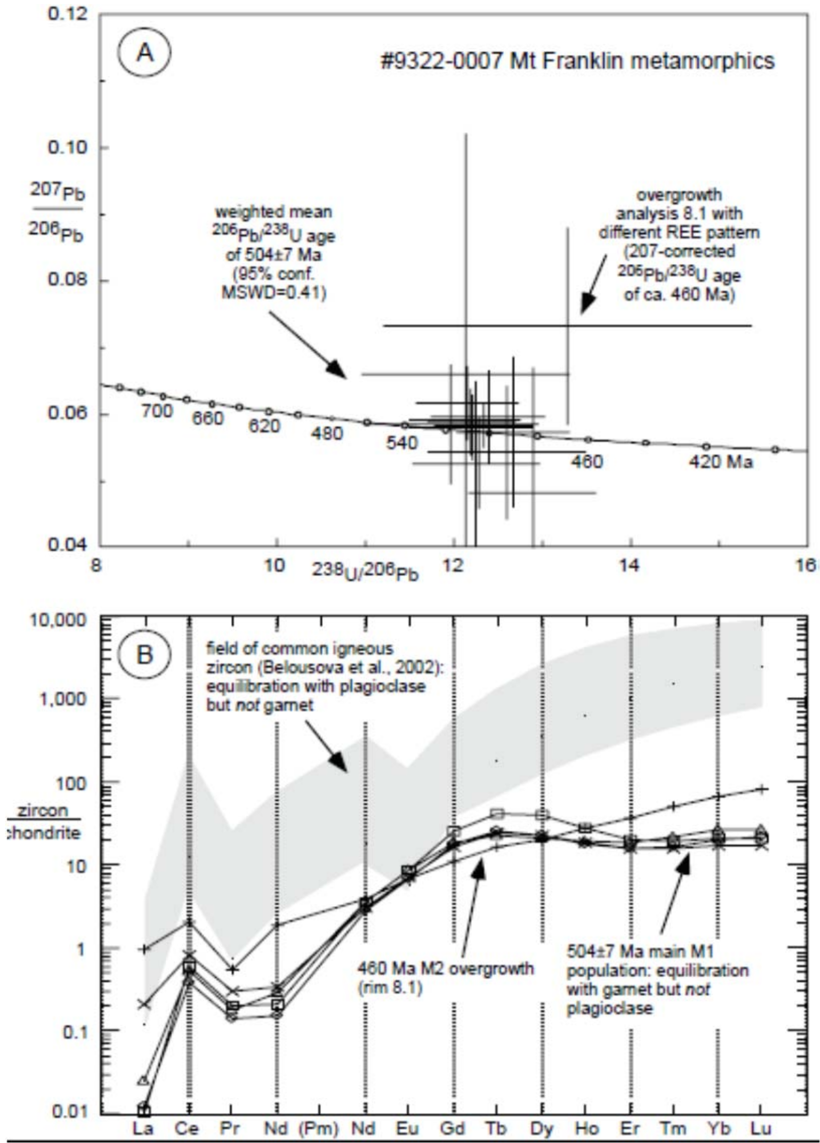

A
#9322-0007 Mt Franklin metamorphics
weighted mean
206Pb/238U age
of 504±7 Ma
(95% conf.
MSWD=0.41)
overgrowth
analysis 8.1 with
different REE pattern
(207-corrected
206Pb/238U age
of ca. 460 Ma)
207Pb/206Pb
238U/206Pb
700
660
620
480
540
460
420 Ma
B
field of common igneous
zircon (Belousova et al., 2002):
equilibration with plagioclase
but not garnet
460 Ma M2 overgrowth
(rim 8.1)
504±7 Ma main M1
population: equilibration
with garnet but not
plagioclase
zircon
chondrite
La
Ce
Pr
Nd
(Pm)
Nd
Eu
Gd
Tb
Dy
Ho
Er
Tm
Yb
Lu

Fig. 4.

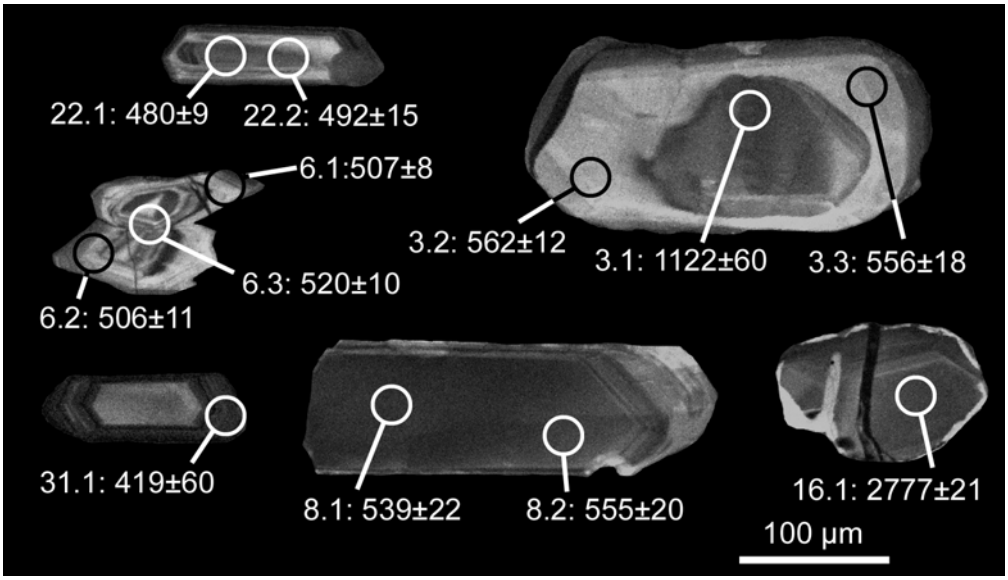

22.1: 480±9
22.2: 492±15
6.1:507±8
6.3: 520±10
6.2: 506±11
3.2: 562±12
3.1: 1122±60
3.3: 556±18
31.1: 419±60
8.1: 539±22
8.2: 555±20
16.1: 2777±21
100 μm

Fig. 5.

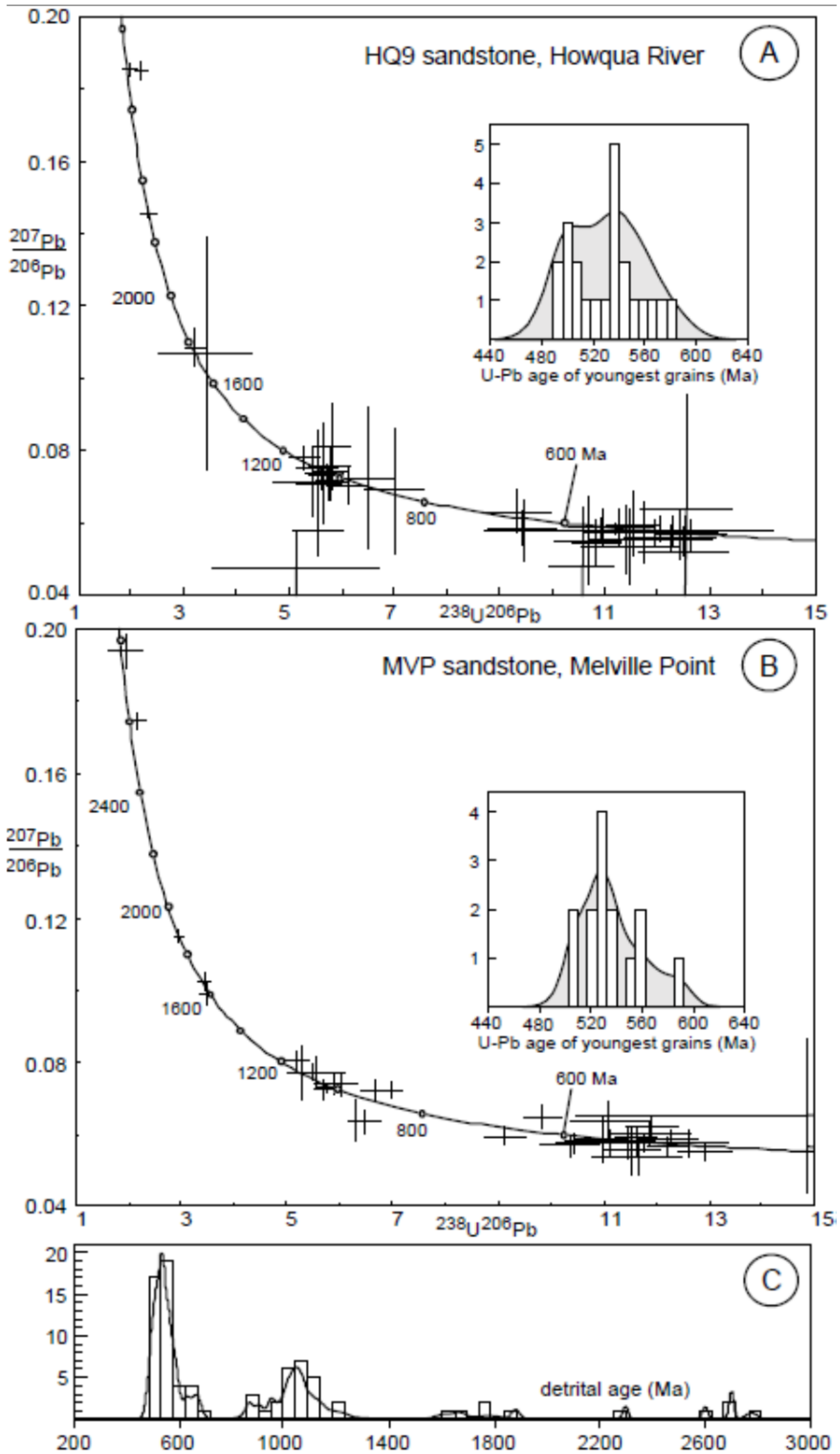
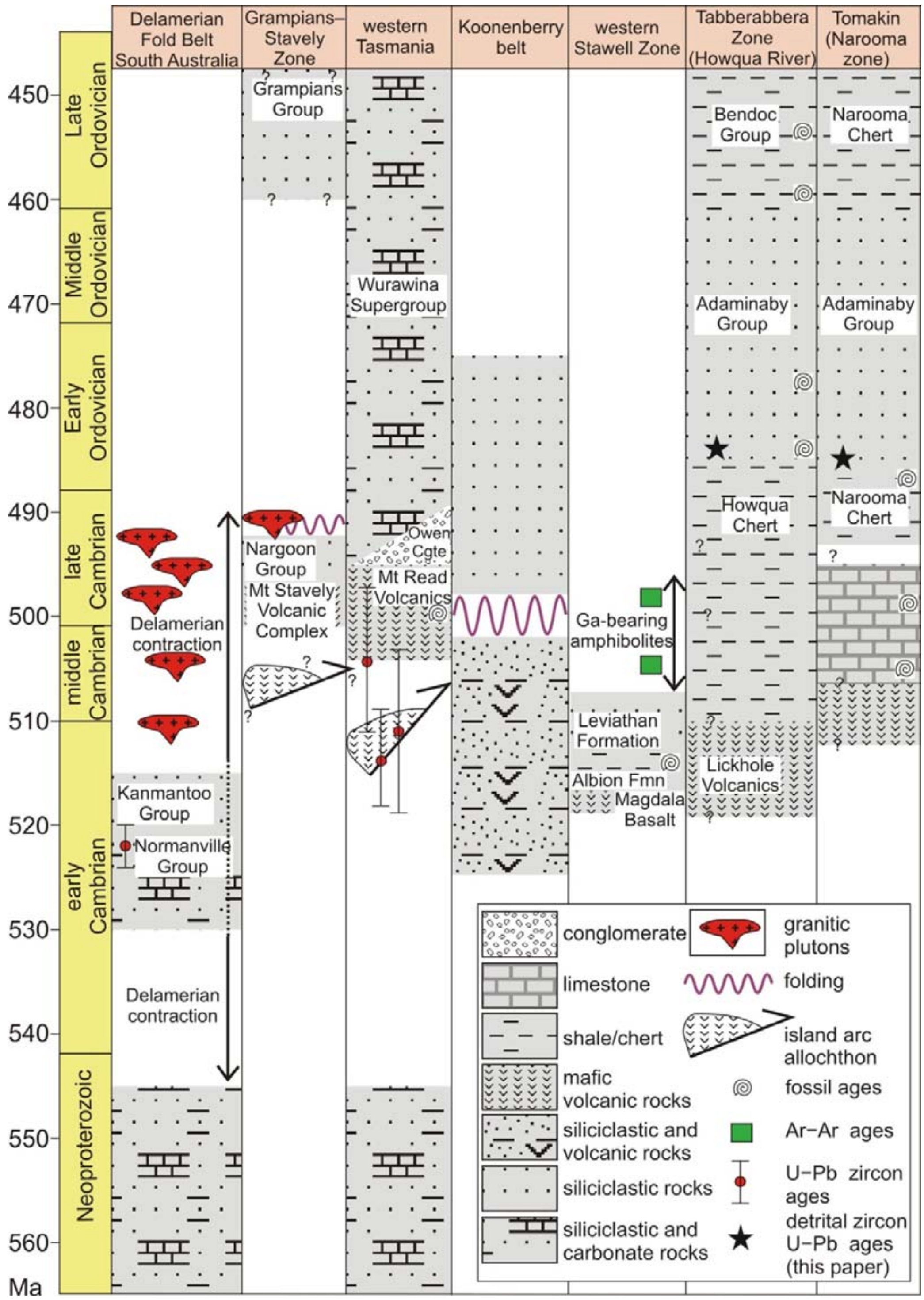
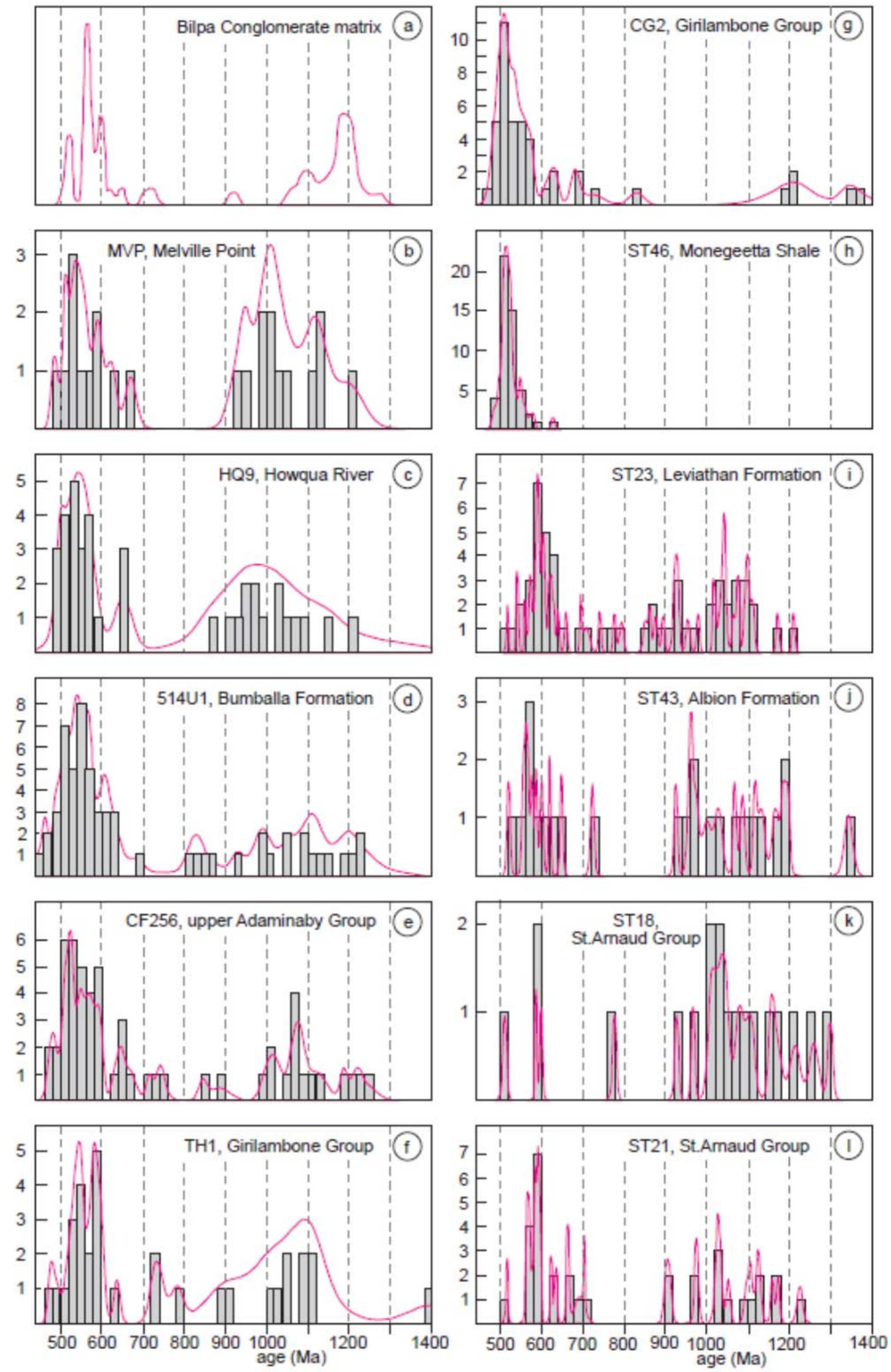

HQ9 sandstone, Howqua River
A
0.20
0.16
0.12
0.08
0.04
207Pb/206Pb
2000
1600
1200
800
600 Ma
1
3
5
7
238U/206Pb
11
13
15
U-Pb age of youngest grains (Ma)
440
480
520
560
600
640
MVP sandstone, Melville Point
B
2400
2000
1600
1200
800
600 Ma
C
detrital age (Ma)
200
600
1000
1400
1800
2200
2600
3000

Fig. 6.



Delamerian Fold Belt South Australia
Grampians–Stavely Zone
western Tasmania
Koonenberry belt
western Stawell Zone
Tabberabbera Zone (Howqua River)
Tomakin (Narooma zone)
Late Ordovician
Middle Ordovician
Early Ordovician
late Cambrian
middle Cambrian
early Cambrian
Neoproterozoic
450
460
470
480
490
500
510
520
530
540
550
560
Ma
Grampians Group
Wurawina Supergroup
Bendoc Group
Narooma Chert
Adaminaby Group
Adaminaby Group
Howqua Chert
Narooma Chert
Owen Cgte
Nargoon Group
Mt Stavely Volcanic Complex
Mt Read Volcanics
Delamerian contraction
Ga-bearing amphibolites
Leviathan Formation
Albion Fmn
Magdala Basalt
Lickhole Volcanics
Kanmantoo Group
Normanville Group
Delamerian contraction
conglomerate
granitic plutons
limestone
folding
shale/chert
island arc allochthon
mafic volcanic rocks
fossil ages
siliciclastic and volcanic rocks
Ar–Ar ages
siliciclastic rocks
U–Pb zircon ages
siliciclastic and carbonate rocks
detrital zircon U–Pb ages (this paper)

Fig. 7.



Bilpa Conglomerate matrix
a
MVP, Melville Point
b
HQ9, Howqua River
c
514U1, Bumballa Formation
d
CF256, upper Adaminaby Group
e
TH1, Girilambone Group
f
CG2, Girilambone Group
g
ST46, Monegeetta Shale
h
ST23, Leviathan Formation
i
ST43, Albion Formation
j
ST18,
St.Arnaud Group
k
ST21, St.Arnaud Group
l
500 600 700 800 900 1000 1200 1400
age (Ma)

Fig. 8.

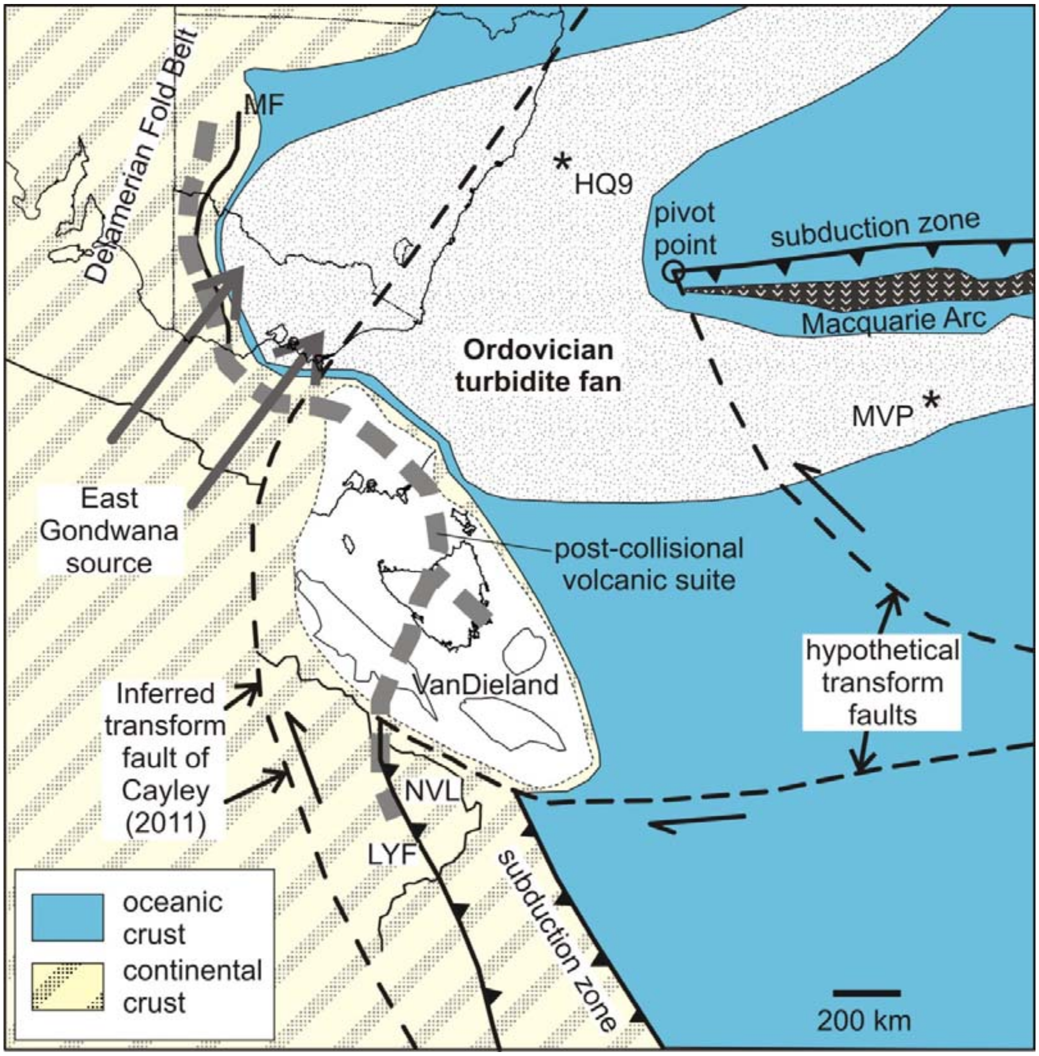

Delamerian Fold Belt
MF
HQ9
pivot point
subduction zone
Macquarie Arc
Ordovician turbidite fan
MVP
East Gondwana source
post-collisional volcanic suite
hypothetical transform faults
VanDieland
Inferred transform fault of Cayley (2011)
NVL
LYF
subduction zone
oceanic crust
continental crust
200 km

Fig. 9.